Ontario

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Open File Report 6140

Geology of the Northern Superior Area, Ontario

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ONTARIO GEOLOGICAL SURVEY

Open File Report 6140

Geology of the Northern Superior Area, Ontario

by

D. Stone

2005

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GEOLOGIC MAP

Preliminary Map P.3545 – Precambrian Geology, Northern Superior Area .... back pocket

*MRD 135 is sold separately from this report

Abstract

The northern Superior area is situated at Sachigo Lake, approximately 700 km north-northwest of Thunder Bay, Ontario. The 14 000 km^2 area represents the northernmost exposure of Archean rocks of the Superior Province in Ontario. Geological mapping of the northern Superior area was completed in consecutive field seasons from 1995 to 2000.

The northern Superior area spans four major east-southeasterly trending greenstone belts interspersed with felsic plutonic domains. From south to north, the greenstone belts include those at Ponask Lake, Stull–Swan lakes, Ellard Lake and Yelling Lake. The area is transected by three regional-scale east-southeasterly trending faults including the Stull–Wunnummin, South Kenyon and North Kenyon faults and associated splays.

New mapping and geochronology have refined the major crustal blocks in the northern Superior area. The Island Lake terrane extends south from the Stull–Wunnumnin fault and includes part of the Stull–Swan lakes greenstone belt and the Ponask Lake greenstone belt as well as intervening plutonic areas. The Island Lake terrane is represented by volcanic sequences of mainly 2.86 to 2.85 Ga and plutonic rocks with an age-range including most of Geon 28. Thorium- and light rare earth element (LREE)-enriched basalt sequences, 3.0 Ga inherited zircon grains and 3.0 Ga Nd model ages for plutons support the interpretation that magmatic rocks of the Island Lake terrane have assimilated older crust. The older crust is possibly part of the 3.0 Ga North Caribou terrane, which is a major crustal block south of the present area.

The Oxford Lake–Stull Lake terrane extends north from the Stull–Wunnummin fault to the South Kenyon fault and includes most of the Stull–Swan lakes greenstone belt, the Ellard Lake greenstone belt and adjacent plutonic areas. Magmatic rocks of the Oxford Lake–Stull Lake terrane are of 2.73 to 2.71 Ga age although older mafic volcanic sequences such as the 2.83 Ga Hayes River assemblage occur locally. The Oxford Lake–Stull Lake terrane represents a mature volcanic arc that developed on or near the margin of the continental Island Lake terrane at 2.73 Ga. Older mafic volcanic sequences were thrust (D_1) onto the continental margin at an early stage in development of the magmatic arc.

The Northern Superior Superterrane extends north from the South Kenyon fault and is composed of felsic plutonic and gneissic rocks as well as supracrustal sequences of the Yelling Lake greenstone belt. Magmatic rocks of the Northern Superior Superterrane have crystallization ages of 2.85 to 2.78 Ga and 2.72 to 2.69 Ga. Inherited zircon grains of 3.57 to 3.21 Ga age and strongly negative ε_{ND} values in plutonic and gneissic rocks suggest that parts of the Northern Superior Superterrane represent a highly assimilated Paleoarchean continental fragment. Supracrustal sequences of the Yelling Lake greenstone belt developed on the continental margin at 2.83 and 2.72 Ga or were tectonically displaced onto the margin after these times.

The three crustal blocks were accreted together from north to south along sutures marked by the South Kenyon and Stull–Wunnummin faults beginning at about 2.71 Ga. The early stages of accretion involved shortening and folding of supracrustal sequences and thrusting on terrane boundary faults (D_2) followed by transcurrent displacement (D_3) on the faults with local development of pull-apart basins wherein late sedimentary sequences developed.

The Ponask Lake greenstone belt is divided into the Sachigo and Pierce assemblages. The base of the Sachigo assemblage is a continental margin-type platform sequence where tonalite cobble conglomerate and komatiite-marble units overly 2.86 Ga tonalite basement. The remainder of the Sachigo assemblage and the Pierce assemblage is composed of mafic to ultramafic volcanic sequences with a lesser component of 2.86 Ga intermediate to felsic volcanic rocks and wacke. The majority of mafic volcanic rocks are enriched in Th and LREE (primitive-mantle-normalized Th>La>Nb). The geochemistry can be interpreted to indicate that the mafic sequences erupted in a volcanic arc or in a continental environment where they were contaminated with Th- and LREE-enriched crustal material. Their association with platform sequences favours the latter of these environments. Slightly depleted basalt sequences (primitive-mantle-normalized Th≈Nb<La) representative of an oceanic environment are also present in the belt. The close spatial association of enriched and depleted basalt sequences can indicate that volcanism occurred in an environment transitional from a continental margin to an ocean basin or that the sequences are tectonically interleaved.

The Stull–Swan lakes greenstone belt is composed of diverse mafic volcanic, intermediate to felsic volcanic and sedimentary assemblages. The Stull assemblage represents variably enriched and depleted basalt sequences south of the Stull–Wunnummin fault and is interpreted to be part of the Island Lake terrane. The Rorke Lake and Edmund Lake assemblages are enriched and depleted basalt sequences, respectively, at the north side of the greenstone belt within the Oxford Lake–Stull Lake terrane. Although undated, these are possibly correlative in age with the 2.83 Ga Hayes River assemblage at Knee Lake, Manitoba. Variably enriched and depleted mafic volcanic greenstone slivers in the Swan Lake area are part of the Swan assemblage.

Central parts of the Stull–Swan lakes greenstone belt at Stull Lake and several greenstone units at Swan Lake are represented by intermediate to felsic, calc-alkalic to alkalic volcanic rocks of the Oxford Lake volcanic assemblage. Primitive-mantle-normalized trace element profiles of the Oxford Lake volcanic assemblage are steeply sloped from left to right due to enrichment in incompatible elements and have troughs for Nb and Ti. These characteristics suggest that the Oxford Lake volcanic assemblage erupted in magmatic arcs. Eruption of the calc-alkalic phases began at 2.73 Ga and was followed by alkalic phases as late as 2.71 Ga. ε_{ND} values for the early calc-alkalic phases are positive indicating that eruption occurred in juvenile and possibly oceanic arcs, whereas ε_{ND} values of the late alkalic phases are negative. The alkalic phases of the Oxford Lake volcanic assemblage probably erupted after terrane accretion in a continental environment where magmas assimilated older LREE-enriched crust.

Wacke sequences of the Oxford Lake sedimentary assemblage are spatially and temporally associated with the Oxford Lake volcanic assemblage and appear to represent marine basin deposits. Coarse conglomerate of the Cross Lake assemblage overlies alkalic phases of the Oxford Lake volcanic assemblage and tends to be concentrated in local basins along major faults. The Cross Lake assemblage contains detrital zircon grains ranging in age from 3.5 to 2.7 Ga, which supports the interpretation that it was deposited after terrane amalgamation and received detritus from the Northern Superior Superterrane as well as the Oxford Lake–Stull Lake terrane.

The Ellard Lake greenstone belt is composed of mafic volcanic sequences of the Rorke Lake and Ellard assemblages. The undated Ellard assemblage has chemically enriched basalt units (Th>La>Nb) as well as depleted basalt and gabbro (Th<Nb<La). A thin, bifurcated unit of the 2.73 Ga Oxford Lake volcanic assemblage and the Oxford Lake sedimentary assemblage occupies the central axis of part of the belt. Conglomerate of the Cross Lake assemblage occurs locally.

The narrow, bifurcated and highly metamorphosed Yelling Lake greenstone belt is composed of several mafic assemblages and small units of the 2.72 Ga Oxford Lake volcanic and sedimentary assemblages. The mafic assemblages include the enriched 2.83 Ga Dadson assemblage (Th=La>Nb) and the depleted 2.72 Ga Yelling assemblage (Th<Nb≈La) and at least one poorly exposed and chemically unknown assemblage. The geochemistry of the Dadson assemblage is interpreted to indicate that it erupted on the margin of the Northern Superior Superterrane at 2.83 Ga or erupted in a volcanic arc and was obducted onto the old continent after that time. In contrast, the Yelling assemblage appears to have developed in an oceanic environment and was faulted into position after 2.72 Ga.

Felsic plutonic rocks are divided into six major suites on the basis of mineralogy, texture and composition. These include the biotite tonalite, gneissic, hornblende tonalite, biotite granite, peraluminous and sanukitoid suites. The biotite tonalite and gneissic suites occur widely in all terranes as plutons and belt-like domains and have crystallization ages ranging from 2.86 to 2.70 Ga. Gneisses of the Northern Superior Superterrane have inherited zircon grains as old as 3.57 Ga. The hornblende tonalite and biotite granite suites also occur widely as elongate plutons and batholiths. Although a biotite granite pluton in the Northern Superior Superterrane has an age of 2.85 Ga, most other intrusions of these suites have ages of 2.72 to 2.69 Ga. Rocks of the peraluminous or two-mica granite suite occur locally as dikes and small plutons cutting migmatized sedimentary units. The sanukitoid suite includes quartz-undersaturated to saturated rocks compositionally variable from diorite through monzonite to monzodiorite and granite. Magmas of the sanukitoid suite were emplaced as oval plutons at 2.71 Ga.

Greenstone sequences are metamorphosed to greenschist and amphibolite facies with the amphibole-bearing assemblages concentrated at greenstone belt margins and in thin greenstone slivers. Zircon overgrowths, which are interpreted to represent metamorphic events, are dated at 2.74 and 2.71 Ga in the Northern Superior Superterrane, 2.72 Ga in the Oxford Lake–Stull Lake terrane and 2.76 Ga in the Island Lake terrane.

Aluminum-in-hornblende barometry indicates that hornblende-bearing plutonic suites (mainly the hornblende tonalite and sanukitoid suites) crystallized at pressures from 9.6 to 1.0 kbars with somewhat higher pressures recorded in the hornblende tonalite suite than in the sanukitoid suite. For the hornblende tonalite suite, the Island Lake terrane is characterized by crystallization pressure of 4 to 6 kbars, whereas two blocks are distinguished on the basis of pressure in the Oxford Lake–Stull Lake terrane: pressure determinations are 6 to 9 kbars in a southern block and 4 to 6 kbars in a northern block. The Northern Superior Superterrane has overall lowest pressures of 2 to 4 kbars. The pressure variations can be explained by dip-slip displacement on boundary faults after intrusion of the hornblende tonalite suite. The Island Lake terrane and Northern Superior Superterrane have underthrust margins of the Oxford Lake–Stull Lake terrane. The northern block of the Oxford Lake–Stull Lake terrane has, in turn, underthrust the southern block of the Oxford Lake–Stull Lake terrane causing it to have overall greatest uplift.

The northern Superior area shows excellent potential for economic concentrations of gold mineralization. Several gold showings including two with substantial inferred reserves are localized along the Stull–Wunnummin fault. This and other regional-scale faults provide first-order exploration targets for gold. At the past-producing Sachigo River Mine, gold is spatially associated with a late mantle-derived pluton of the sanukitoid suite. The sanukitoid plutons may have been a source of metal or a heat-engine for driving hydrothermal cells leading to deposition of gold in nearby structural traps such as faults. Deformed greenstone and plutonic sequences adjacent to sanukitoid plutons are recommended for gold exploration.

The sand in beaches of the northern Superior area has anomalous concentrations of kimberlite indicator minerals including pyrope, picro ilmenite, chrome diopside, chromite and olivine compared to beaches in other parts of the western Superior Province. Although the source of the kimberlite indicator minerals is unknown, they provide an indication that kimberlite with possible associated diamond can occur in the area.

Indications of base metal mineralization are identified at several localities in the northern Superior area. Gossan zones, one of which has elevated values of copper and nickel, occur in mafic to ultramafic sequences of the Ponask Lake greenstone belt. Base metal occurrences comprising disseminated sphalerite, galena and chalcopyrite are associated with mafic to felsic volcanic sequences at Ponask Lake and with the alkalic diorite intrusion at Gilleran Lake. Several showings at Gilleran Lake have been explored previously.

Rare metals including Li, Rb, Cs, Be, Nb, Ta and Ga are noted in highly fractionated dikes of the peraluminous granite suite in adjacent parts of Manitoba, particularly near major faults. Peraluminous granite dikes localized along the major faults in the present area as well as the "Carb" Lake carbonatite intrusion provide exploration targets for rare metals.

Location, Geologic Setting, Project Scope

The northern Superior area is situated at Sachigo Lake approximately 700 km north-northwest of Thunder Bay, Ontario. The mapped area (Figure 1) spans 14 000 km^2 of sparsely inhabited land and is bounded on the south and north approximately by latitudes 54° and 55° and to the east and west by longitude 91° and the Manitoba–Ontario provincial border, respectively.

The northern Superior area transects the northernmost Superior Province of Ontario. The area is underlain by a series of Archean greenstone belts and plutonic domains onlapped to the north by Paleozoic sedimentary rocks of the Hudson Bay Lowlands.

Owing largely to remoteness and poor exposure, northern parts of the Superior Province have had relatively less bedrock mapping, geologic research and mineral exploration than more accessible areas to the south. The present survey was initiated to characterize the geology, mineral potential and tectonic setting of the northern Superior area and to improve the overall level of geologic information in the north. The bedrock mapping has defined four major greenstone belts including the Ponask Lake, Stull–Swan lakes, Ellard Lake and Yelling Lake greenstone belts (Figure 2) as well as six suites of plutonic rocks. Several mineral showings and indications of gold, base metal and diamond mineralization have been identified. Although early tectonic subdivisions of the Superior Province (e.g., Card and Ciesielski 1986) showed this area to be part of the broad Sachigo Subprovince, the present work in combination with collaborative projects by the geological surveys of Manitoba and Canada have redefined major crustal blocks in the north. These include the Island Lake terrane, Oxford Lake–Stull Lake terrane and the Northern Superior Superterrane (Figure 3).

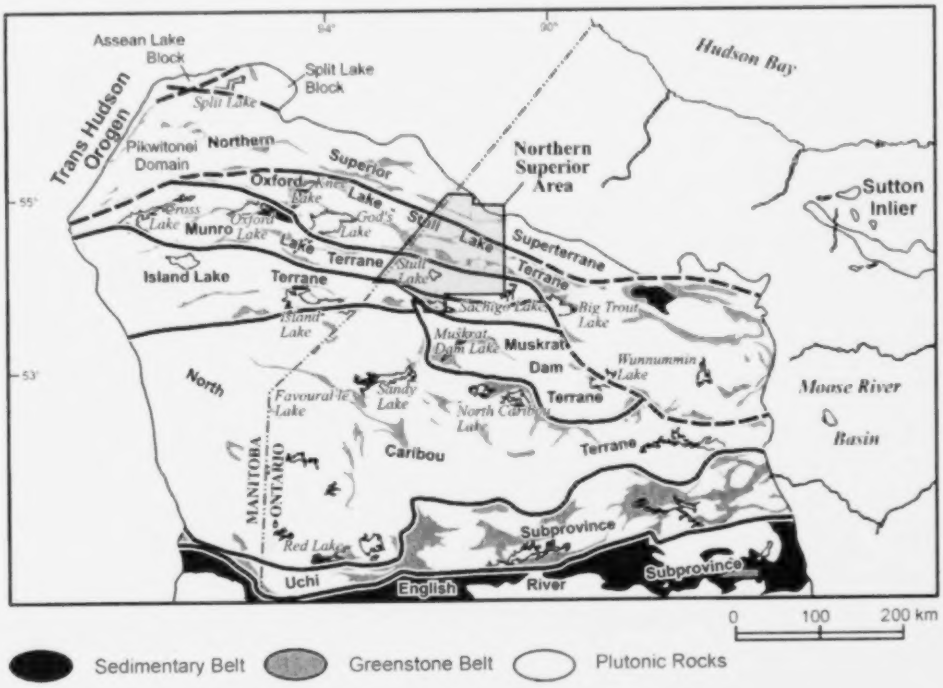
94°
90°
55°
53°
Assean Lake Block
Split Lake Block
Split Lake
Trans Hudson Orogen
Hudson Bay
Pikwitonei Domain
Northern Superior Superterrane
Northern Superior Area
Sutton Inlier
Oxford Lake Stull Lake Terrane
Knee Lake
Cross Lake
Oxford Lake
God's Lake
Munro Lake Terrane
Island Lake Terrane
Stull Lake
Island Lake
Sachigo Lake
Big Trout Lake
Muskrat Dam Lake
Muskrat Dam Terrane
Wunnummin Lake
North Caribou Terrane
Favourable Lake
Sandy Lake
North Caribou Lake
Moose River Basin
MANITOBA
ONTARIO
Red Lake
Uchi Subprovince
English River Subprovince
0
100
200 km
Sedimentary Belt
Greenstone Belt
Plutonic Rocks

Figure 1. Terrane subdivisions of the northwest Superior Province from Thurston et al. (1991) and Skulski et al. (1999).

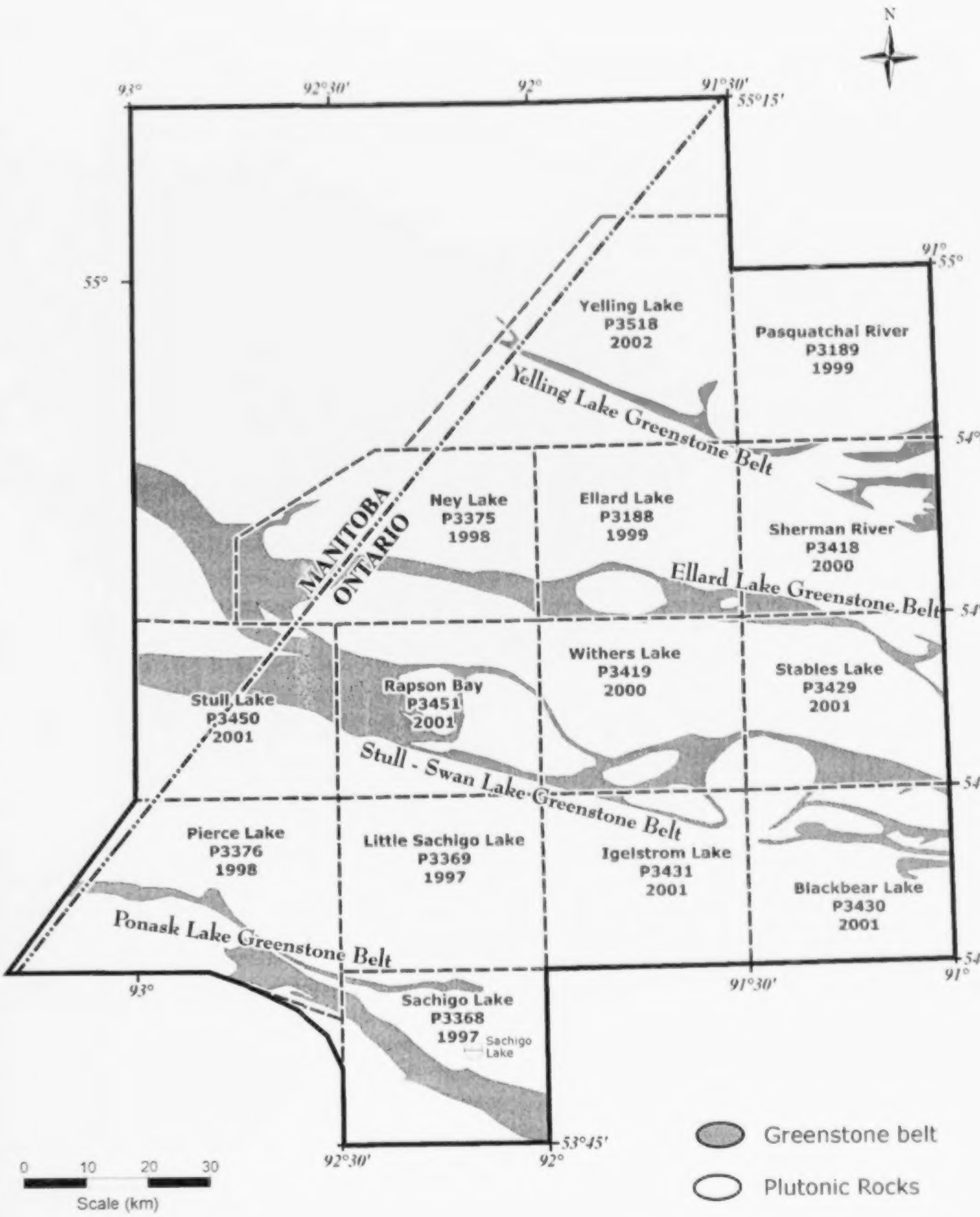
N
93°
92°30'
92°
91°30'
55°15'
55°
91°
55°
54°45'
54°30'
54°15'
54°
91°
93°
91°30'
53°45'
92°30'
92°
MANITOBA
ONTARIO
Yelling Lake
P3518
2002
Pasquatchai River
P3189
1999
Yelling Lake Greenstone Belt
Ney Lake
P3375
1998
Ellard Lake
P3188
1999
Sherman River
P3418
2000
Ellard Lake Greenstone Belt
Withers Lake
P3419
2000
Stables Lake
P3429
2001
Rapson Bay
P3451
2001
Stull Lake
P3450
2001
Stull - Swan Lake Greenstone Belt
Pierce Lake
P3376
1998
Little Sachigo Lake
P3369
1997
Igelstrom Lake
P3431
2001
Blackbear Lake
P3430
2001
Ponask Lake Greenstone Belt
Sachigo Lake
P3368
1997
Sachigo Lake
0
10
20
30
Scale (km)
Greenstone belt
Plutonic Rocks

Figure 2. Major greenstone belts of the northern Superior area and index to published geologic maps.

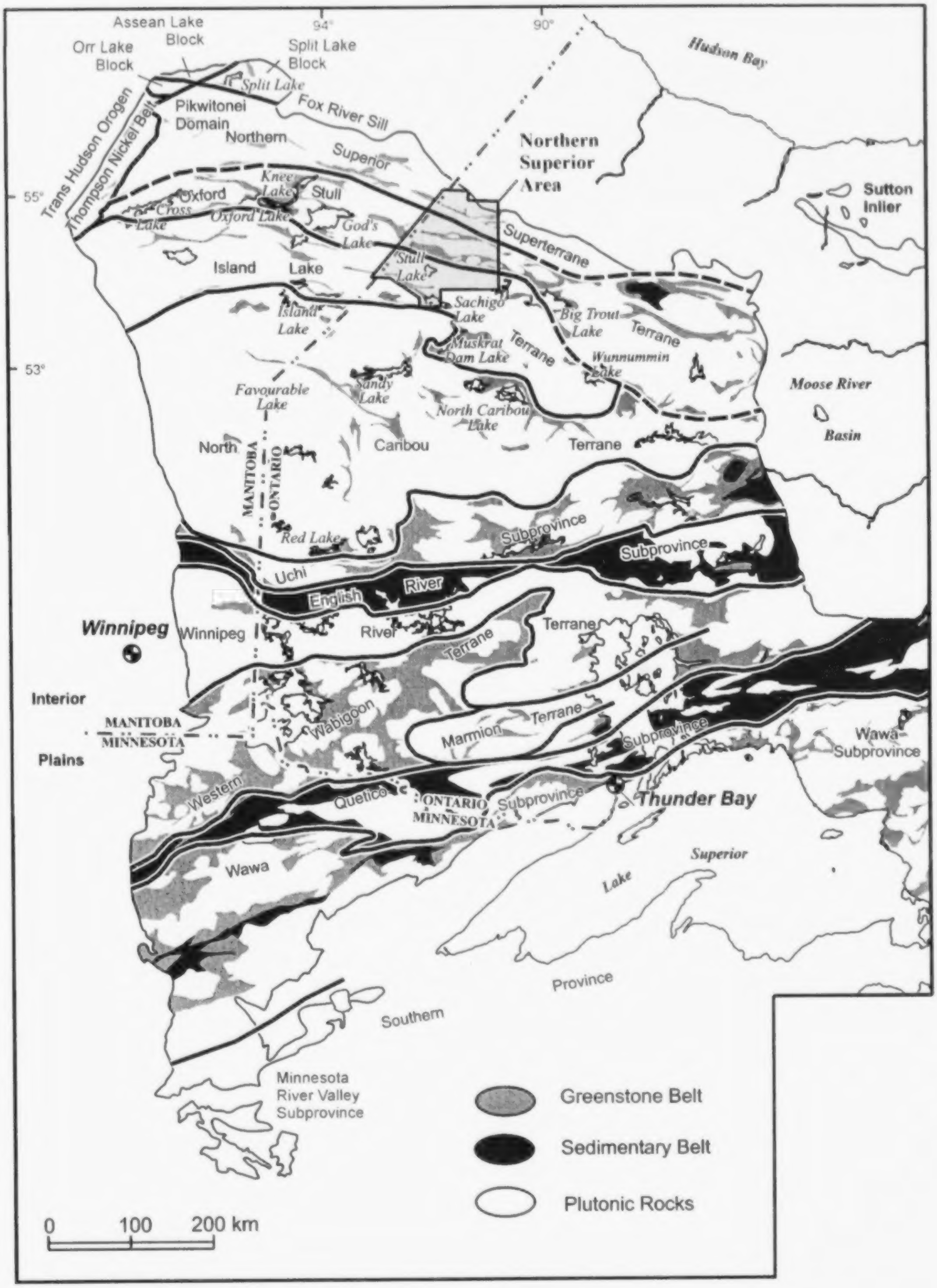
Assean Lake Block
94°
90°
Hudson Bay
Orr Lake Block
Split Lake Block
Split Lake
Fox River Sill
Trans Hudson Orogen
Thompson Nickel Belt
Pikwitonei Domain
Northern
Superior
Northern Superior Area
55°
Knee Lake
Oxford
Stull
Cross Lake
Oxford Lake
God's Lake
Superterrane
Sutton Inlier
Island
Lake
Stull Lake
Sachigo Lake
Island Lake
Big Trout Lake
Terrane
Muskrat Dam Lake
Terrane
Wunnummin Lake
53°
Favourable Lake
Sandy Lake
North Caribou Lake
Moose River
Basin
North
Caribou
Terrane
MANITOBA
ONTARIO
Red Lake
Subprovince
Subprovince
Uchi
English
River
Winnipeg
Winnipeg
River
Terrane
Terrane
Interior
Plains
MANITOBA
MINNESOTA
Wabigoon
Marmion
Terrane
Subprovince
Wawa Subprovince
Western
Quetico
ONTARIO
MINNESOTA
Subprovince
Thunder Bay
Superior
Lake
Wawa
Province
Southern
Minnesota River Valley Subprovince
Greenstone Belt
Sedimentary Belt
Plutonic Rocks
0
100
200 km

Figure 3. Subprovinces and terranes of the western Superior Province, *after* Thurston et al. (1991), Böhm et al. (2003) and Tomlinson et al. (2004).

Fieldwork

Bedrock of the northern Superior area was mapped at a scale of 1:50 000 in the summers of 1995 to 2000. Airphotographs taken between 1954 and 1976 and supplied by the National Airphoto Library were used for navigation and as a plotting base. Geologic information was captured digitally in the field using AutoCAD® version 12 and subsequently AutoCAD® version 14 coupled with Fieldlog® version 3.2. Fourteen preliminary geologic maps were produced in subsequent seasons (*see* Figure 2) using 1:50 000 scale base maps of the National Topographic System in Zone 15, NAD 27. These maps were subsequently compiled into the 1:250 000 scale geologic map that accompanies this report.

In addition to bedrock mapping, samples of till and modern alluvium were collected from the northern Superior area and processed for indicator grains of gold, metamorphosed or magmatic sulphide minerals and kimberlite minerals. Stone (2000b, 2001) discusses the results of this work.

Fieldwork was done mainly from a series of tent camps established on lakes through the area and re-supplied by float-equipped aircraft from Red Sucker Lake, Manitoba. Helicopters were used extensively to transport mapping teams to remote area.

Acknowledgements

Independent mapping was done by P. Pufahl, J. Carter, J. Hallé, M. Lange and E. Chaloux in various field seasons. Junior assistants were O. Kreuzer, O. Welte, D. Kendall, K. Schendel, L. Hubbard, C. Miller, E. Lecuyer, J. Dougherty, C. Klages, D. Hardy, J. Bjorkman, J. Arnold, K. Bjorkman, L. Lepage, N. Petersen and D. Quick.

T. Corkery and P. Thurston assisted with geologic interpretations in the field. J. Ketchum, formerly of the Jack Satterly Geochronology Laboratory, Toronto greatly improved the geologic interpretation of the area by determining the ages of rocks. Good logistical support was provided by E. Cull of Red Sucker Air Services. The manuscript benefited from comments by J. Parker. The diagrams were drawn by L. Roy and J. Chartrand.

Unless otherwise noted, all data for geochemical analyses presented in this report were provided by the Geoscience Laboratories, Ontario Geoservices Centre in Sudbury.

Topography, Drainage, Forests

Topography of the northern Superior area is extremely subdued and slopes gently to the north. Elevations above sea level range from 248 m at Sachigo Lake in the southern part of the area to 134 m at Yelling Lake in the north. The Sachigo interlobate moraine (Wetiko Hills) extends northerly through the area and comprises the highest topographic feature, locally reaching 50 m above the surrounding land. Topography of most of the area is marked by a series of low (up to 10 m high), southwesterly oriented ridges and valleys interspersed with elongate lakes and muskeg swamps. The linear landforms appear to represent drumlinized till plains.

Drainage is northward through the Stull, Echoing and Sachigo rivers. These rivers merge with the God's River and Severn River and flow to Hudson Bay.

Muskeg swamps and boreal forest cover the northern Superior area. Broad, poorly drained areas support sparse, stunted black spruce and tamarack giving way to black spruce and jack pine forests on drier, sandy soil. White birch and poplar grow on shorelines and on silty soil representing glaciolacustrine deposits. Well-drained areas including the crests of drumlins and the Sachigo interlobate moraine can support commercially valuable forest resources although these areas tend to be ravaged by forest fires and are variably regenerated by thick young trees.

Previous Mapping and Mineral Exploration

Downie (1937) and Satterly (1937) mapped parts of the Stull–Swan lakes greenstone belt in the area of the Manitoba–Ontario border. Also in 1937, V. B. Meen traversed some major waterways in the area and made reconnaissance maps of parts of the Ponask Lake, Ellard Lake and Yelling Lake greenstone belts (Meen 1937). Bennett and Riley (1969) provided the first systematic mapping at a scale of 1 inch to 2 miles for the northern Superior area. Brief investigations of the Ponask Lake, Stull Lake and Little Stull Lake (Manitoba) areas were made by Thurston, Cortis and Chivers (1987), Atkinson, Parker and Storey (1990) and Corkery (1981), respectively.

Prospectors were active in the 1930s and made several discoveries, the most notable of which was high-grade gold mineralization in a narrow vein at Foster Lake in the eastern Ellard Lake greenstone belt. The Sachigo River Mine operated at Foster Lake from 1938 to 1941 and produced 1 634 826 g of gold and 190 543 g of silver from 71 995 t of ore (Edwards 1944). This represents the only mineral production from the area.

Greenstone belts of the northern Superior area were explored systematically in the 1970s for base metals and gold by several major companies including Sherritt Gordon Mines Limited, Great Plains Development Company of Canada Limited and the International Nickel Company (Assessment Files, Resident Geologist's Office, Red Lake). Atkinson, Parker and Storey (1990) provide a summary of mineral exploration in the western Stull–Swan lakes greenstone belt.

In recent decades, Westmin Resources Limited and Noranda Exploration Company Limited and subsequently Wolfden Resources Limited with Bema Gold Corporation have outlined substantial gold reserves at Little Stull Lake and Twin Lakes, Manitoba (Richardson et al. 1996). The gold mineralization occurs within sheared volcanic rocks of the western Stull–Swan lakes greenstone belt.

The Current Work as Part of a Collaborative Project

The northern Superior project was done collaboratively with Operation Superior of Manitoba Energy and Mines and the Western Superior NATMAP Project of the Geological Survey of Canada. Operation Superior was focussed on production of maps and interpretations of large greenstone belts in northern Manitoba including western extensions of the Ellard Lake and Stull–Swan lakes greenstone belts (Little Stull Lake area) in Manitoba (Corkery, Skulski and Whalen 1997; Corkery and Skulski 1998; and Corkery et al. 2000). Cross-border mapping exercises and field trips have helped to correlate geologic units as well as the terminology of geologic units between Manitoba and Ontario (compare Corkery et al. 2000b and Stone, Hallé and Pufahl 2001b).

Media including rock, till, B-horizon soil, humus and vegetation were sampled over the Edmund Lake and Sharpe Lake greenstone belts on the Manitoba side of the provincial border and chemically analysed (Fedikow et al. 1998) as part of Operation Superior. The multimedia survey also included identification of kimberlite indicator minerals and provides a set of data in Manitoba comparable to the kimberlite dataset in the northern Superior area of Ontario (Stone 2001).

Geologists of the Geological Survey of Canada assisted with mapping in the Little Stull Lake area of Manitoba and did age determinations and Nd isotopic studies on plutonic and supracrustal rocks in Manitoba as well as the present area. The results, summarized by Skulski et al. (2000) were critical to identification of crustal blocks and definition of the tectonic evolution of the northern Superior area. Stone et al. (2004) and Corkery and Stone (in press) produced 1:250 000 scale compilation maps of geology and tectonostratigraphic assemblages in the northern Superior area and adjacent parts of Manitoba. These maps are part of a series of NATMAP compilation maps from the western Superior Province.

Previous Work in the Far-Northwestern Superior Province: An Overview

SUBDIVISIONS OF SUPRACRUSTAL AND PLUTONIC ROCKS: MANITOBA

The terminology and subdivisions of supracrustal rocks of the northwestern Superior Province was established and evolved mainly due to work on greenstone sequences in northern Manitoba (*see* Figure 1). Subdivision of the Archean greenstone strata was initiated by Wright (1931) who applied the terms Hayes River Group and Oxford Lake Group to sequences of predominantly volcanic and sedimentary rocks, respectively, at Oxford Lake, Manitoba. Downie (1937) subsequently adopted these group names and succinctly divided the supracrustal strata at Stull Lake into the Hayes River and Oxford Lake groups. The Hayes River Group represents an older sequence of predominantly mafic volcanic lavas, minor intermediate to felsic volcanic rocks and sediments, whereas the Oxford Lake Group included an overlying sequence of conglomerate, arkose and greywacke.

Since the 1930s, rocks comprising northern Manitoba greenstone belts have been widely subdivided into Hayes River and Oxford Lake groups (e.g., Gilbert 1985). Greenstone belts of Ontario underwent comparable subdivisions into the Keewatin Type (predominantly mafic volcanic lithologies) and Timiskaming Type (late clastic sedimentary and felsic volcanic rocks). These early subdivisions of greenstone strata can be examined with respect to the Stull Lake area by comparing the maps of Downie (1937) and Satterly (1937).

Based on fieldwork in 1973, Hubregtse (1985) determined that fragmental volcanic rocks previously assigned to the Hayes River Group at Oxford Lake were part of the Oxford Lake Group and went on to divide the Oxford Lake Group into a lower volcanic subgroup and upper sedimentary subgroup. Geochemical work of Brooks et al. (1982) showed that the intermediate to felsic commonly fragmental volcanic rocks of the Oxford Lake volcanic subgroup have calc-alkalic to alkalic affinity. At the type localities on Oxford Lake and Knee Lake the Oxford Lake sedimentary subgroup is composed of conglomerate (with a high proportion of felsic volcanic and tonalite clasts), arkose and greywacke deposited in fluvial to deep marine basin environments (Hubregtse 1985; Syme et al. 1997, 1998, 1999). Unconformities were observed locally where the Hayes River Group is overlain by the Oxford Lake volcanic subgroup, which is in turn overlain by the Oxford Lake sedimentary subgroup.

The application of geochronology since the 1980s confirmed depositional time-gaps associated with the unconformities and has established the ages of the Hayes River and Oxford Lake groups. Felsic volcanic strata associated with the mafic lavas of the Hayes River Group have been dated at 2834 to 2827 Ma at Knee Lake (Nos. 1, 5, 6, 7 of Table 1). The youngest detrital zircon grain in greywacke associated with volcanic rocks of the Hayes River Group at Knee Lake is 2822 Ma (Nos. 8, 9 of Table 1) and indicates that some sedimentary sequences in northern Superior greenstone belts could have developed at the time of Hayes River volcanism. At Island Lake, Stevenson and Turek (1992) and Corfu and Lin (2000) obtained ages of 2861 to 2852 Ma (Nos. 2, 3 of Table 1) representing mafic sequences, which they included with the Hayes River Group. Recent work by Parks et al. (2002, 2003) has subdivided the Hayes River Group at Island Lake into 5 geologically and geochemically distinct assemblages with eruption ages at ca. 2890, 2850 and 2744 Ma. The eruption age for the oldest assemblage is inferred from detrital zircon grains in a sedimentary lens within mafic volcanic sequences.

Plutonic rocks are not extensively dated in northern Manitoba. The available data indicates several tonalite, tonalite gneiss and gabbro bodies in the range of 2883 to 2807 Ma as well as other plutonic rocks of post 2750 Ma age (Nos. 10 to 13 of Table 1 and references therein). The 2883 to 2807 Ma plutonic rocks occur south of Knee Lake, south of Cross Lake and at Island Lake, whereas post 2750 Ma plutonic rocks occur widely. Corkery and Stone (in press) reported ages of 2785 and 2812 Ma on biotite tonalite and tonalite gneiss at Red Cross Lake.

Volcanic rocks and associated intrusions of the Oxford Lake volcanic subgroup show a range of ages from 2734 to 2717 Ma (Nos. 17 to 21 of Table 1) at Knee Lake, Margaret Lake and Little Stull Lake. Deposition of the Oxford Lake sedimentary subgroup is constrained by the youngest detrital zircon grains to be younger than 2706 Ma at Oxford Lake, 2713 Ma at Little Stull Lake and 2707 Ma at Knee Lake (Nos. 22, 23, 26, 27 of Table 1).

With the aid of geochronology, further subdivisions have been made of supracrustal sequences in northern Manitoba greenstone belts. For example, Corkery, Davis and Lenton (1992) established 3 supracrustal sequences at Cross Lake. These include the Pipestone Lake Group (2758 Ma; No. 14 of Table 1), a thick sequence of mainly pillowed basalt, which is unconformably overlain by clastic sediments and interbedded rhyodacite of the Gunpoint Group (2729 Ma; No. 15 of Table 1). The Gunpoint Group is overlain by the Cross Lake Group (<2713 Ma; No. 16 of Table 1) that comprises a sequence of basal conglomerate fining upward through sandstone and siltstone with shoshonitic basalt flows at the top.

At Island Lake, Godard (1963a,b) identified the sedimentary Island Lake Group unconformably overlying mafic volcanic sequences. Corfu and Lin (2000) showed that the Island Lake Group was deposited over a protracted time interval. The basal greywacke contains detrital zircon grains with ages in the range of 2896 to 2821 Ma (No. 24 of Table 1) and is cut by a 2744 Ma tonalite pluton indicating that deposition occurred in the 2821 to 2744 Ma interval. In contrast, greywacke in upper sections of the Island Lake Group were deposited after 2712 Ma, the age of the youngest detrital zircon grain (No. 25 of Table 1). Parks et al. (2002) reported zircon grains as old as 3014 Ma in the Island Lake Group and suggested that deposition probably occurred in the interval of 2712 to 2699 Ma, the latter of which is the age of a crosscutting dike dated by Turek et al. (1986).

The volcanic and sedimentary subgroups of the Oxford Lake Group have come under scrutiny in recent years by federal and Manitoba geologists. Syme et al. (1998) confirmed the interpretation of Hubregtse (1985) that the Oxford Lake volcanic subgroup represents a late sequence of predominantly fragmental volcanic rocks of calc-alkalic to alkalic composition erupted from 2.73 to 2.71 Ga, however evidence of complexities within the subgroup has been found. The geochemical and isotopic work of Skulski et al. (2000) indicates that rocks of the Oxford Lake volcanic subgroup have evolved from early calc-alkalic sequences erupted in oceanic arcs to late alkalic sequences developed in continental arcs and pull-apart basins. This implies two possible parts of the volcanic subgroup.

Table 1. Selected U-Pb age determinations, northern Manitoba.

No	Group/Complex (dated rock)	Age (Ma)	Area	Reference
1	Hayes River	≈2830	Knee Lake	D. Davis cited in Syme et al. 1997
2	Hayes River or Jubilee?	2852±1.5	Island Lake	Corfu and Lin (2000)
3	Hayes River or Jubilee?	2861±12	Island Lake	Stevenson and Turek 1992
4	*Hayes River or Jubilee? (wacke)*	*2858±2 to 2847±2*	*Island Lake*	*Corfu and Lin (2000)*
5	Hayes River (lapilli tuff)	2834+3-2	Knee Lake	Corkery et al. (2000a)
6	Hayes River (tuff)	2832±2	Knee Lake	Corkery et al. (2000a)
7	Hayes River (rhyolite)	2827+5-4	Knee Lake	Corkery et al. (2000a)
8	*Hayes River (wacke)*	*2853±10 to 2829±9*	*Knee Lake*	*Corkery et al. (2000a)*
9	*Opischikona sediments*	*2973±12 to 2822+18-15*	*Knee Lake*	*Corkery et al. (2000a)*
10	Felsic gneiss complex	2883	South of Knee Lake	D. Davis cited in Corkery et al. (2000a)
11	Clearwater Bay tonalite	2839±3	Southern Cross Lake	Corkery and Lenton (1989)
12	Tonalite	2825±2	Island Lake	Corfu and Lin (2000)
13	Gabbro	2807±1	Island Lake	Corfu and Lin (2000)
14	Pipestone Lake (gabbro pegmatite)	2758±3	Cross Lake	Corkery, Davis and Lenton (1992)
15	Gunpoint (felsic volcanic)	2729±2	Cross Lake	Corkery, Davis and Lenton (1992)
16	*Cross Lake (sandstone)*	*3547±2 to 2713±3*	*Cross Lake*	*Corkery, Davis and Lenton (1992)*
17	Oxford Volcanic	2726±2	Little Stull Lake	Corkery and Skulski (1998)
18	Oxford Volcanic (synvolcanic tonalite)	2734±2	Margaret Lake	Corkery and Heaman (1998)
19	Oxford Volcanic (quartz+feldspar porphyry)	2717±3	Little Stull Lake	Davis and Moore (1991)
20	Oxford Volcanic (dacite tuff)	2722±3	Knee Lake	Corkery et al. (2000a)
21	Oxford Volcanic	2719.9±1.4	Knee Lake	D. Davis cited in Corkery et al. (2000a)
22	*Oxford Sedimentary*	*2706+4-2*	*Oxford Lake*	*D. Davis cited in Syme et al. (1997)*
23	*Cross Lake*	*2905±8 to 2713±5*	*Little Stull Lake*	*Skulski et al. (2000)*
24	*Island Lake (basal quartz-rich sequence)*	*2896.1±2.1 to 2821.1±7*	*Island Lake*	*Corfu and Lin (2000)*
25	*Island Lake (upper sequence)*	*2938.6±8.1 to 2712±2*	*Island Lake*	*Corfu and Lin (2000)*
26	*Oxford Sedimentary (wacke)*	*2798±12 to 2707+9-8*	*Knee Lake*	*Corkery et al. (2000a)*
27	*Oxford sedimentary or Cross Lake? (Conglomerate)*	*3647±2 to 2711±2*	*Knee Lake*	*D. Davis cited in Corkery et al. (2000a)*
28	(layered tonalite gneiss)	3540	Assean Lake	Böhm et al. (2000)
29	*(wacke)*	*3749±1*	*Assean Lake*	*Böhm et al. (2000)*
30	*(wacke)*	*3330 to 3050*	*Assean Lake*	*Böhm et al. (2000)*

Ages shown in italics are obtained from detrital zircon grains in sedimentary rocks

Late clastic sedimentary sequences in northern greenstone belts have been widely assigned to the Oxford Lake sedimentary subgroup. Recent work by Manitoba geologists (T. Corkery, Geologist, Manitoba Energy and Mines, personal communication, 2002) shows however that the Oxford Lake sedimentary subgroup at its type locality on Oxford Lake is not representative of most other late sedimentary sequences in northern greenstone belts. The Oxford Lake sedimentary subgroup at Oxford Lake has a high proportion of volcanic clasts and was deposited mainly in a subaqueous basin environment. In contrast, late sedimentary sequences such as at Little Stull Lake comprise arkose, arkosic sandstone and conglomerate and are interpreted to have been laid down mainly in fluvial-alluvial environments (Corkery, Skulski and Whalen 1997). T. Corkery (Geologist, Manitoba Energy and Mines, personal communication, 2002) suggests that rocks of the Cross Lake Group (Corkery, Davis and Lenton 1992) provide a well-described analogue to late fluvial-alluvial sequences in northern greenstone belts.

As a further refinement of terminology, Manitoba geologists have redefined various lithologic Groups and Subgroups as assemblages of supracrustal rocks. Hence, the Hayes River Group is known as the Hayes River assemblage and the Oxford Lake volcanic and sedimentary subgroups are known as the Oxford Lake volcanic and sedimentary assemblages. The Cross Lake Group is called the Cross Lake assemblage. Recent work (discussed further below) shows significant variations in the age and composition of mafic sequences formerly assigned to the Hayes River Group or Hayes River assemblage. Hence, the Hayes River assemblage is generally restricted to its type locality at Knee Lake and other assemblage names are used for mafic sequences elsewhere. Following these revisions, Beaumont-Smith et al. (2003), Corkery and Stone (in press) and Stone et al. (2004) produced maps in which mafic sequences formerly assigned to the Hayes River Group are subdivided into assemblages such as the Stull, Edmund Lake and Rorke Lake assemblages. Late sedimentary sequences deposited in marine basins are included with the Oxford Lake sedimentary assemblage and fluvial-alluvial sequences are part of the Cross Lake assemblage.[1] Broadly, the Oxford Lake sedimentary assemblage is interpreted to have been deposited in marine basins associated with volcanic arcs of the early calc-alkalic phases of the Oxford Lake volcanic assemblage. The Cross Lake assemblage is spatially associated with and overlies the late alkalic phase of the Oxford Lake volcanic assemblage.

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[1] The Hayes River and Oxford Lake groups are subdivided into assemblages on the basis of age and depositional environment. Not all assemblages are listed for the Hayes River Group.

Age (Ga)	Group	Assemblage (area)	Depositional Environment(s)
2.83	Hayes River	Hayes River (Knee Lake)	Volcanic arc or continental margin; oceanic
?		Stull (Stull Lake)	Volcanic arc or continental margin; oceanic
?		Rorke Lake (Stull Lake-Little Stull Lake)	Volcanic arc or continental margin
?		Edmund Lake (Edmund Lake to Stull Lake)	Oceanic
2.85		Jubilee (Island Lake)	Continental margin, arc
2.73 to 2.72	Oxford Lake volcanic subgroup	Early calc-alkalic phase of the Oxford Lake volcanic assemblage (Oxford Lake-Stull Lake terrane)	Mature island arc
2.71		Late alkalic phase of the Oxford Lake volcanic assemblage (Stull Lake)	Continental arc
2.73 to 2.71	Oxford Lake sedimentary subgroup	Oxford Lake sedimentary assemblage (Oxford Lake-Stull Lake terrane)	Marine basins associated with the calc-alkalic phase of the Oxford Lake volcanic assemblage
<2.71		Cross Lake assemblage (Oxford Lake-Stull Lake terrane)	Fluvial-alluvial fans locally associated with alkalic phase of Oxford Lake volcanic assemblage

SUBDIVISIONS OF SUPRACRUSTAL AND PLUTONIC ROCKS: FAR-NORTHWESTERN ONTARIO

Most major greenstone belts in the far-northwestern Superior Province of Ontario have been mapped (e.g. Satterly 1938; Hudec 1964; Bennett and Riley 1969; Ayres 1969; Thurston, Sage and Siragusa 1979; and Wilson 1987) although few mapping surveys were accompanied by geochemical and geochronological research. Supracrustal strata have been subdivided into Keewatin and Timiskaming Types (Satterly 1938) or more commonly into mafic, intermediate and felsic volcanic rocks and clastic and chemical sedimentary rocks without the assignment of group names such as in Manitoba. Thurston, Osmani and Stone (1991) provided an overview of the geology of northern greenstone belts and initiated the subdivision of greenstone belts in this part of Ontario into supracrustal assemblages on the basis of available geology, geochemistry and geochronology. A notable feature of greenstone belts in the far-northwestern Superior Province is that they contain a mix of Neoarchean[2] and Mesoarchean assemblages in contrast with predominantly Neoarchean assemblages in the southern Superior Province (Corfu and Davis 1992).

The Favourable Lake greenstone belt represents one of the first greenstone belts to be subdivided on the basis of detailed geochronology (Ayres and Corfu 1991; Corfu and Ayres 1991). These authors showed that the Favourable Lake greenstone belt is composed of five assemblages of predominantly volcanic rocks ranging in age from 2924 to 2725 Ma (Nos. 1 to 5 of Table 2). Similarly, the geochronologic work of Corfu and Wood (1986) showed that the North Spirit Lake greenstone belt is a collage of volcanic and sedimentary assemblages of divergent age (Nos. 7 to 11 of Table 2). These include 3023 Ma volcanic arc sequences of the North Spirit assemblage, mixed volcanic and sedimentary rocks of the Disrupted assemblage (2820 to 2950 Ma) and quartz arenite of the Nemakwis assemblage that has zircon grains whose ages cluster at 2986 Ma. The North Spirit Lake greenstone belt also contains Neoarchean volcanic assemblages including the 2735 Ma Hewitt assemblage and the calc-alkalic to alkalic Bijou Point assemblage dated at 2731 Ma.

At most localities the margins of the Favourable Lake and North Spirit greenstone belts are marked by intrusive Neoarchean felsic plutonic rocks although scattered remnants of older crust are identified. The latter include a 2950 Ma tonalite enclave at Favourable Lake and a 2925 Ma biotite tonalite pluton at North Spirit Lake (Nos. 6 and 12 of Table 2 and references therein). Corfu and Ayres (1991) and Corfu and Wood (1986) identified a 2960 to 3000 Ma granodiorite clast in the North Trout assemblage and a 3001 Ma tonalite clast in sedimentary sequences of the North Spirit greenstone belt, respectively.

The Sandy Lake greenstone belt has a spectrum of Mesoarchean and Neoarchean assemblages (Nos. 13 to 17 of Table 2). At Sandy Lake, the 2945 Ma volcanic North Sandy assemblage is in fault contact

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[2]

Summary of time-terms used in this report			
Age (billions of years before present)	Eon	Era	Geon
3.3	Archean	Paleoarchean	33
3.2			32
3.1		Mesoarchean	31
3.0			30
2.9			29
2.8			28
2.7		Neoarchean	27
2.6			26
2.5			25
2.4	Proterozoic	Paleoproterozoic	24

with the sedimentary West Arm and Keewaywin assemblages. These latter quartz-rich sedimentary assemblages contain detrital zircon grains ranging in age from 3022 to 2895 Ma but lack Neoarchean grains suggesting that deposition of the sedimentary sequences occurred in Mesoarchean time. The Sandy Lake greenstone belt also contains a small unnamed volcanic assemblage dated at 2734 Ma and is intruded by Neoarchean dikes and plutons (Davis and Moore 1991; Davis and Stott 2001).

Supracrustal sequences of the North Caribou Lake greenstone belt developed episodically at about 3.0 Ga and 2.87 Ga. Included in the older category are the volcanic Agutua Arm and South Rim assemblages that occur in the western and southern parts of the belt and are dated at 2981 Ma (Nos. 18 and 19 of Table 2). Quartz arenite platform sequences of the Keeyask assemblage unconformably overlie the Agutua Arm assemblage and contain a single population of 2979 Ma zircon grains (No 21 of Table 2). The Keeyask assemblage is cut by a quartz porphyry dike, which has an age of 2863 Ma and constrains deposition of the sediments to the 2979 to 2863 Ma interval. Sandstone sequences of the Eyapamikama assemblage, which occupy the central axis of the North Caribou Lake greenstone belt, contain zircon populations with ages of 2980 and 2962 Ma (No. 22 of Table 2). This, combined with the lack of Neoarchean detrital grains, suggests that the Eyapamikama assemblage developed during the early stage of greenstone belt evolution.

The second stage in development of the North Caribou Lake greenstone belt is represented by the North Rim volcanic assemblage, which occurs along the north side of the greenstone belt and contains a variety of zircon grains. These include inherited zircon grains comparable in age to older volcanic assemblage of the North Caribou Lake greenstone belt and 2870 Ma zircon grains that are interpreted to represent the age of eruption and crystallization of the North Rim assemblage (No. 20 of Table 2; Davis and Stott 2001). Quartz arenite sequences in the Heaton assemblage of the eastern North Caribou Lake greenstone belt contain detrital zircons ranging in age from 2980 to 2853 Ma, the latter of which represents the maximum age of deposition for the sequence (No. 23 of Table 2).

Felsic plutonic rocks intruded margins of the North Caribou Lake greenstone belt at approximately the same time as supracrustal sequences developed within the belt. The early episode of plutonism is represented by the Weagamow Lake batholith, which intruded the western end of the North Caribou Lake greenstone belt at 2990 Ma (No. 27 of Table 2). At the second stage of plutonism, the De Bliquey Lake tonalite gneiss was emplaced north of the belt at 2856 Ma (No. 24 of Table 2). Several phases of the North Caribou Lake batholith that range in age from 2870 to 2864 Ma intruded to the south (Nos. 25 and 26 of Table 2). Granodiorite of the North Caribou Lake batholith contains inherited zircons as old as 3017 Ma, probably indicating that the magma erupted through and assimilated older crust.

Sporadic geochronology has been done at other localities in far-northwestern Ontario (Nos. 28 to 30 of Table 2). For example, a 2852 million-year-old zircon grain is derived from sandstone sequences at Wunnummin Lake and trondhjemite at Windigo Lake (50 km southwest of North Caribou Lake) contains a spectrum of 3037 to 2804 Ma inherited grains. In contrast, two volcanic sequences in the Muskrat Dam greenstone belt are 2734 Ma old. Broadly, these data record protracted although somewhat episodic magmatic and sedimentary activity from 3.0 to 2.7 Ga in the far-northwestern Superior Province of Ontario.

Table 2. Selected U/Pb age determinations, far-northwestern Ontario.

No	Assemblage/Suite* (dated rock)	Age (Ma)	Area	Reference
1	Setting Net (intermediate flow)	2924±1.5	Favourable Lake	Corfu and Ayers (1991)
2	South Trout (diorite with intermediate volcanic and sedimentary rocks)	2870+8/-2	Favourable Lake	Corfu and Ayres (1991)
3	East Trout (felsic tuff with mafic flows)	2858+5/-4	Favourable Lake	Corfu and Ayres (1991)
4	Northwind (tuff, felsic pyroclastic rocks)	2734 with inherited grains to 2930	Favourable Lake	Corfu and Ayres (1991)
5	North Trout (felsic tuff with greywacke, conglomerate and mafic flows)	2725±2	Favourable Lake	Corfu and Ayres (1991)
6	Tonalite enclave	2950±5	Favourable Lake	Corfu, Krogh and Ayres (1985)
7	North Spirit (tuff breccia with mafic to felsic volcanic rocks)	3023±2	North Spirit Lake	Corfu and Wood (1986)
8	Disrupted (lapilli tuff)	2820 to 2950	North Spirit Lake	Corfu and Wood (1986)
9	*Nemakwis (quartz arenite)*	*2986+3/-2*	*North Spirit Lake*	*Corfu and Wood (1986)*
10	Hewitt (crystal tuff with mafic to felsic volcanic and sedimentary rocks)	2735±10 with 2862 xenocrysts	North Spirit Lake	Corfu and Wood (1986)
11	Bijou Point (quartz diorite with calc-alkalic to alkalic volcanic rocks)	2731±2	North Spirit Lake	Corfu and Wood (1986)
12	Tonalite pluton	2925+10/-8	North Spirit Lake	Stevenson (1995)
13	North Sandy (felsic and mafic volcanic rocks)	2945	Sandy Lake	D. Davis cited in Thurston, Osmani and Stone (1991)
14	*Keewaywin (pebbly arkose)*	*3017±1 to 2904±2*	*Sandy Lake*	*Davis and Stott (2001)*
15	*West Arm (Chert clast conglomerate)*	*3022±3 to 2895±11*	*Sandy Lake*	*Davis and Moore (1991)*
16	*Keewaywin (Quartz arenite)*	*2968±9 to 2959±3*	*Sandy Lake*	*Davis and Moore (1991)*
17	Unnamed (Ash flow)	2734±3	Sandy Lake	Davis and Moore (1991)
18	Agutua Arm (rhyolite with mafic volcanic rocks)	2981±1	North Caribou Lake	Davis and Moore (1991)
19	South Rim (rhyolite tuff)	2981.9±0.8	North Caribou Lake	Davis and Stott (2001)
20	North Rim (rhyolite tuff)	2870±2; inherited grains to 2980±2	North Caribou Lake	Davis and Stott (2001)
21	*Keeyask (quartz arenite)*	*2979±1*	*North Caribou Lake*	*Davis and Moore (1991)*
22	*Eyapamikama (sandstone)*	*2980.3±1.1 to 2962.6±1.3*	*North Caribou Lake*	*Davis and Stott (2001)*
23	*Heaton (quartz arenite)*	*2980.8±0.8 to 2853±1*	*North Caribou Lake*	*Davis and Stott (2001)*
24	Tonalite gneiss, De Bliquey Lake	2856±3	North Caribou Lake	Davis and Moore (1991)
25	Tonalite, North Caribou batholith	2870±2	North Caribou Lake	Davis and Moore (1991)
26	Biotite granodiorite, North Caribou batholith	2864±1.1; inherited grains to 3017±1	North Caribou Lake	Davis and Stott (2001)
27	Tonalite, Weagamow batholith	2990±1	North Caribou Lake	Davis and Moore (1991)
28	Trondhjemite, Shinbone Lake	Inherited grains 3037±2 to 2804±2	Windigo Lake	Davis and Stott (2001)
29	*Sandstone*	*2852±2*	*Wunnummin Lake*	*Davis and Moore (1991)*
30	Muskrat Dam (rhyolite)	2734	Muskrat Dam Lake	D. Davis in Thurston, Osmani and Stone (1991)

* Assemblage names are from Thurston, Osmani and Stone (1991)
Ages shown in italics are obtained from detrital grains in sedimentary rocks

Table 3. U-Pb age data for volcanic, plutonic and sedimentary units of the northern Superior area.

Map No.	Sample No.	Area	UTM east	UTM north	Rock Type	Greenstone belt or area	Assemblage or plutonic suite	Crystallization Age	Maximum Depositional Age	Inherited Age	Metamorphic Age	Reference
Island Lake terrane												
1	00DS418	Ponask Lake	522729	5985714	conglomerate	Ponask Lake	Sachigo		2863.8+/-0.7			7
2	96DS315	E. of Pierce Lake	513200	5992150	biotite granite	Ponask Lake	Biotite granite	2723+/-6				1
3**	WX97M13	Richardson Arm	520000	6022200	hornblende+biotite tonalite gneiss	S of Stull-Swan lakes	Gneissic	2848+/-7				1
4	96DS95	N. of Little Sachigo	560500	6007700	biotite tonalite	S of Stull-Swan lakes	Biotite tonalite	2855+/-5			2758	1
5	00DS420	Ponask Lake	522279	5986136	biotite tonalite	Ponask Lake	Biotite tonalite	2863.3+/-0.7				7
6	00DS421	Pierce Lake	503621	5995464	quartz-rich wacke	Ponask Lake	Pierce		2863.2+/-0.7			7
7**	PT87-1	Ponask Lake	522800	5985700	quartz-rich conglomerate	Ponask Lake	Sachigo		2865+/-1			6
8**	PT87-3	Ponask Lake	531000	5978500	quartz porphyry in mafic volcanic	Ponask Lake	Sachigo	2857+/-2				6
Oxford Lake-Stull Lake terrane												
10**	GRS88-523	Little Stull Lake	518950	6047600	quartz+feldspar porphyry	Stull-Swan lakes	Oxford Lake volcanic (1)	2717+/-3				6
11	88GRS501	S of Ellard Lake	572790	6040110	fragmental felsic volcanic	Ellard Lake	Oxford Lake volcanic (1)	2732.4+/-1.0				2
12	00DS424	Stull Lake	528046	6029975	quartz+felspar porphyritic rhyolite	Stull-Swan lakes	Oxford Lake volcanic (1)	2722.5+6.1/-4.4				4
13	00DS427	E side of Stull L.	527699	6031622	fragmental felsic volcanic	Stull-Swan lakes	Oxford Lake volcanic (1)	2718.2+/-1.0				4
14	00DS422a	Rapson Bay	534158	6029777	felsic tuff	Stull-Swan lakes	Oxford Lake volcanic (1)	2720.9+/-0.9				4
15	00DS422b	Rapson Bay	534158	6029777	sandstone	Stull-Swan lakes	Oxford Lake sedimentary		2718.7+/-2			4
16	00DS426	E Side of Stull L.	526790	6026431	sandstone	Stull-Swan lakes	Oxford Lake sedimentary		2722.6+/-2.7			4
17	00DS425	E Side of Stull L.	528230	6028447	sandstone	Stull-Swan lakes	Cross Lake		2709.4+/-1.7			4
18**	22	Little Stull Lake	524000	6043900	arkose	Stull-Swan lakes	Cross Lake		2713+/-5			1
19**		Little Stull Lake	520000	6046950	aphyric dacite tuff	Stull-Swan lakes	Oxford Lake volcanic (1)	2726+/-2		2742, 2752		3
20**	WX97M5	Kistigan Lake	527600	6050700	hornblende granodiorite	N of Stull-Swan lakes	Hornblende tonalite	2715+/-8		2747		1
21**	SNB97536	Rorke Lake	531400	6042500	hornblende granodiorite	N of Stull-Swan lakes	Sanukitoid	2710+/-6				1
22	00DS405	Thorne River	591889	6043889	hornblende granodiorite	N of Ellard Lake	Hornblende tonalite	2718.7+/-1.5				7
23**	DDH 3, 4	Carb Lake	563600	6069000	carbonatite	N of Ellard Lake	Carbonatite	1822+/-96*				5
24	00DS406	E. of Lacey Lake	585400	6038450	biotite tonalite	S of Ellard Lake	Biotite tonalite	2733.7+/-1.7		2849+/-12		7
25	00DS403b	Foster Lake	603224	6041246	quartz+feldspar porphyry	Ellard Lake	Oxford Lake volcanic (1)	2730.4+/-1.3		2735.6+/-2.2		7
26	00DS412	Swan Lake	613029	6013912	plagioclase porphyry	Stull-Swan lakes	Oxford Lake volcanic (1)	2724.6+/-1.0			2715.1+/1.5	7
Northern Superior superterrane												
30	00DS407	Dadson Lake	628792	6059701	rhyodacite tuff	Yelling Lake	Dadson	2838+/-1.1		2846.6+/-1.3		4
31	00DS410	W. of Dadson L.	622893	6060628	felsic tuff	Yelling Lake	Oxford Lake volcanic (1)	2718.1+/-.9				4
32	00DS415	E. of Yelling Lake	591299	6075055	gabbro	Yelling Lake	Mafic/ultramafic	2716+/-1.3				4
33	97DS157	E. of Yelling Lake	595700	6077700	hornblende granodiorite-monzonite	N of Yelling Lake	Sanukitoid	2714+/-8		2813, 2720		1
34	88GRS503	Yelling Lake	574910	6077790	biotite granodiorite	N of Yelling Lake	Biotite tonalite	2722.5+/-2				2
35	00DS409	NW of Dadson L.	613385	6068798	hornblende granodiorite	N of Yelling Lake	Hornblende tonalite	2716.2+/-1.2				7
36	00DS417	Echoing River	573673	6073188	biotite tonalite	S of Yelling Lake	Biotite tonalite	2822.0+/-0.8				7
37	97DS65	N of Yelling L.	569300	6090850	biotite granodiorite	N of Yelling Lake	Biotite granite	2846+/-5		2872	2717	1
38	97DS59	Yelling Lake	576200	6085300	hornblende+biotite tonalite gneiss	N of Yelling Lake	Gneissic	2814+/-4		3209, 3572	2741	1
39	00DS84b	N Yelling Lake	580050	6101000	granite dike	N of Yelling Lake	Biotite granite	2690-2711				4
40	00DS84	N Yelling Lake	580050	6101000	biotite tonalite gneiss	N of Yelling Lake	Gneissic	-		3292, 2854		7
41	00DS414	E Yelling Lake	587071	6081396	foliated tonalite	N of Yelling Lake	Gneissic	2701.1+4.0/-1.8		2745, 2840		7
42	00DS416	E Yelling Lake	591584	6072722	intermediate tuff	Yelling Lake	Oxford Lake volcanic (1)	2718.1+2/-1.8				7

References

1	Skulski et al. (2000)	5	Sage (1987)
2	Davis and Stott (2001)	6	Davis and Moore (1991)
3	Corkery and Skulski (1998)	7	Ketchum (2002)
4	Ketchum and Davis (2001)		

* K-Ar biotite method

** Locations approximate

(1) early calc-alkalic phase of the Oxford Lake volcanic assemblage

UTM coordinates are in meters for Zone 15, NAD 27

Terrane Subdivisions of the Far-Northwestern Superior Province: A Review

Early tectonic analyses summarized by Card and Ciesielski (1986) have shown two major subdivisions including the Sachigo Subprovince and the Pikwitonei granulite domain in the far-northwestern Superior Province of Ontario and Manitoba. Thurston, Osmani and Stone (1991) undertook further subdivision of the Sachigo Subprovince into five major crustal blocks. These authors identified crustal blocks on the basis of common depositional environments of greenstone sequences and common duration of geologic events within the blocks. At most localities, block boundaries were defined on the basis of faults inferred from airborne geophysical magnetic data and an accretionary model was proposed for the assembly of the crustal blocks.

The North Caribou terrane is a major element of the subdivided Sachigo Subprovince (*see* Figure 1). The North Caribou terrane is characterized by Mesoarchean quartz-rich sedimentary sequences (platform sequences of Thurston and Chivers 1990) and a spectrum of 3.0 to 2.9 billion year old volcanic and plutonic rocks. It is a large crustal block including parts of the Favourable Lake, Sandy Lake and North Caribou Lake greenstone belts and is regarded as a continental nucleus or central block against which other terranes were accreted southward and northward (Thurston, Osmani and Stone 1991).

The Island Lake terrane was defined by Thurston, Osmani and Stone (1991) to include 2.88 Ga volcanic sequences of the Hayes River Group at Island Lake and magmatic rocks of similar age extending along the north margin of the North Caribou terrane. In comparison with rocks of the North Caribou terrane, the Island Lake terrane had younger volcanic and plutonic rocks and was thought to have fewer quartz-rich platform sequences. The Island Lake terrane was interpreted to have accreted to the North Caribou terrane at 2.73 to 2.72 Ga because it lies outboard of the 2.73 Ga Muskrat Dam terrane.

Thurston, Osmani and Stone (1991) defined the Muskrat Dam terrane on the basis of 2734 Ma volcanic sequences in the Muskrat Dam greenstone belt and aeromagnetic patterns east of the belt that are characteristic of unmetamorphosed plutonic rocks. The Muskrat Dam terrane was thought to represent a Neoarchean crustal fragment within a collage of older blocks.

The Munro Lake terrane of Thurston, Osmani and Stone (1991) is an east-southeasterly-elongated crustal element north of the Island Lake terrane (*see* Figure 1). Greenstone belts of the Munro Lake terrane are composed primarily of mafic lavas with calc-silicate and quartz-rich sediments and few rhyolites. Mafic lavas of the 2.76 Ga Pipestone Lake Group at Cross Lake, Manitoba (Corkery, Davis and Lenton 1992) are representative of the greenstone sequences. The Munro Lake terrane was interpreted to have cratonized at about 2.74 Ga and docked with the continental mass to the south at 2.72 to 2.71 Ga (Thurston, Osmani and Stone 1991).

The Oxford Lake–Stull Lake terrane of Thurston, Osmani and Stone (1991) encompasses an extensive area of the northern Superior Province in Manitoba and Ontario and includes major greenstone belts such as at Oxford Lake, Knee Lake and Stull Lake. The greenstone belts are composed of 2.83 Ga basalts of the Hayes River Group and 2.73 to 2.71 Ga volcanic and sedimentary sequences of the Oxford Lake Group. The Oxford Lake–Stull Lake terrane was interpreted to be a circa 2.86 Ga arc sequence intruded by plutonic rocks up to 2.73 Ga and subsequently affected by wrench faulting and development of the Oxford Lake Group in pull-apart basins from 2.73 to 2.71 Ga. Accretion to the continental mass to the south occurred at 2.71 Ga.

Since the work of Thurston, Osmani and Stone (1991), extensive geologic research and geochronology has indicated evidence of pre 2.9 Ga crust in northern Manitoba. This includes highly metamorphosed gneisses in the Pikwitonei and Split Lake blocks (Krogh et al. 1986; Böhm 1998; Heaman, Böhm and Creaser 1998; Böhm et al. 1999). Further, a Paleoarchean crustal block containing 3540 Ma tonalite gneiss and greywacke with 3.1 to 3.7 Ga detrital zircon grains has been identified at Assean Lake near Split Lake (Nos. 28 to 30 of Table 1 and Böhm et al. 2000). Although the Assean Lake block is interpreted to have been tectonically joined to the Superior craton after 2.7 Ga (Böhm et al. 2003) the Pikwitonei and Split Lake blocks appear to have been involved in the pre 2.70 Ga accretion of the Superior Province as discussed above. Skulski et al. (1999) drew attention to widely distributed evidence on the basis of inherited zircon ages and Nd isotopic systematics for pre 2.9 Ga crust at the north margin of the Superior Province. Areas of highly assimilated and potentially ancient crust include the Pikwitonei and Split Lake blocks, northern parts of the present area (Nos. 38 and 40 of Table 3) and the Minto Block of northern Quebec (Percival et al. 1992). Skulski et al. (1999, 2000) defined the Northern Superior Superterrane as a highly reworked Paleoarchean crustal block lying north of the Oxford Lake–Stull Lake terrane (*see* Figure 1) and joined to the growing Superior craton after 2.71 Ga.

Towards a New Tectonic Subdivision of the Far-Northwestern Superior Province

The collaborative research by federal and provincial surveys as well as universities has greatly added to the geologic database of the far-northwestern Superior Province in recent years. New geochronology, geochemistry and mapping has aided in refining the age and depositional environment of supracrustal sequences as well as providing a basis for regional correlation of rock units. This new information facilitates refined terrane analysis (*see* Figure 3).

ISLAND LAKE TERRANE

At Island Lake, basalt sequences previously included with the Hayes River Group have been subdivided into 5, fault-bounded assemblages. The Savage Island shear zone, is a large west-northwest-trending boundary fault interpreted to have been active at 2.70 Ga based on the age of a partly deformed crosscutting dike (Parks et al. 2003). The Jubilee assemblage occurs south of the Savage Island shear zone and is enriched in Th, variably enriched in light rare earth elements (LREE) and is depleted in Nb, Ta and Ti relative to other primitive-mantle-normalized trace elements (Parks et al. 2002). The Jubilee assemblage has been dated at several localities and has ages in the range of 2.86 to 2.85 Ga. It is intruded by 2.83 and 2.78 Ga tonalite plutons. The volcanic assemblages north of the Savage Island shear zone have broadly similar geochemistry to the Jubilee assemblage and show a range of ages. These include the Whiteway assemblage (2.89 Ga), Pipe Point assemblage (2.85 Ga) and the Loonfoot assemblage (2.74 Ga) and the undated York assemblage (Parks et al. 2002, 2003).

The Th- and LREE-enriched geochemistry of volcanic assemblages at Island Lake provides an indication that the volcanic magmas assimilated older crustal material or else erupted in volcanic arcs (Pearce 1996). The close association of the Island Lake assemblages (particularly the Jubilee assemblage) with the 3.0 Ga North Caribou terrane suggests that the mafic magmas could have passed through and assimilated older crustal material. Hence, the Th- and LREE-enriched geochemistry of these assemblages possibly reflects that their magmas were emplaced into a continental margin where contamination occurred. In contrast, basalt sequences formerly assigned to the Hayes River Group at Knee Lake, Edmund Lake and northern Stull Lake have mainly flat primitive-mantle-normalized trace element profiles indicative of eruption in oceanic environments (Syme et al. 1999; Corkery and Heaman

1998; Corkery and Skulski 1998) although enriched varieties occur locally. Further, the age of volcanism for the Hayes River assemblage in its type locality at Knee Lake is 2.83 Ga. These data indicate that volcanic assemblages at Island Lake could have erupted in a different environment about 20 Ma earlier than the majority of mafic sequences located north in the Oxford Lake–Stull Lake terrane.

The contaminated mafic volcanic sequences of the Jubilee assemblage at Island Lake are broadly similar in age and depositional environment to certain volcanic strata at Ponask Lake (data of this study discussed later) and in the North Rim assemblage at North Caribou Lake. Ages of these sequences are 2.87 to 2.85 Ga and are shown by analyses Nos. 2 and 3 of Table 1, No. 20 of Table 2 and No. 8 of Table 3. Although geochemical data is not available for the North Rim assemblage at North Caribou Lake, this assemblage contains 3.0 Ga inherited zircon grains indicating that the magma has passed through and assimilated older crustal material.

In Manitoba, critical tonalite plutons were emplaced south of Cross Lake, south of Knee Lake and at Island Lake from 2.88 to 2.83 Ga (Nos. 11, 10 and 12 of Table 1). In Ontario, biotite tonalite complexes intruded the Ponask Lake greenstone belt and south margin of the Stull–Swan lakes greenstone belt at 2.86 to 2.85 Ga (Nos. 3, 4 and 5 of Table 3). At North Caribou Lake, the 2.87 to 2.86 Ga North Caribou batholith and DeBliquey Lake gneiss complex were emplaced on either side of the North Caribou Lake greenstone belt. The DeBliquey Lake gneisses occur within the Muskrat Dam terrane and indicate that older plutonic rocks are more extensive than was envisaged by Thurston, Osmani and Stone (1991). Collectively, these data show that extensive mafic to felsic magmatism occurred along the north margin of the North Caribou terrane at 2.88 to 2.83 Ga and mainly from 2.87 to 2.85 Ga.

At North Caribou Lake, the close structural position of the North Rim volcanic assemblage to 3.0 Ga sequences combined with 3.0 Ga inherited zircon grains strongly suggests that these mafic magmas erupted through the margin of the North Caribou craton at 2.87 Ga. Also, by virtue of 3.0 Ga detrital zircon grains, the North Caribou continent is implicated as a source for detritus in sedimentary basins represented by the Heaton assemblage at North Caribou Lake (No 23 of Table 2) and the Island Lake assemblage at Island Lake (Parks et al. 2003). The tonalite plutons between the Ponask Lake and Stull–Swan lakes greenstone belts have negative ε_{ND} values and 3.0 Ga model ages (Skulski et al. 2000), which support involvement of 3.0 Ga crustal material in the emplacement of these magmas at 2.86 to 2.85 Ga.

Evidence cited above supports the interpretation that magmatic rocks distributed widely from Island Lake to North Caribou Lake and as far north as Stull Lake have interacted with older crust comparable in age to the North Caribou terrane. It is unclear however whether the older crust represents the North Caribou terrane or alternatively several 3.0 Ga crustal blocks that have been subsequently joined. Some data from north of Island Lake and North Caribou Lake favours the latter of these options. For example, at Ponask Lake the majority of mafic volcanic sequences are enriched in certain trace elements characteristic of crustal contamination, however depleted basalts representative of an ocean basin environment are also present in the greenstone strata (*see* "Ponask Lake Greenstone Belt"). Further, in the Muskrat Dam greenstone belt, 2.73 Ga volcanic arc sequences of the Muskrat Dam assemblage are correlative with the calc-alkalic phase of the Oxford Lake volcanic assemblage. Skulski et al. (2000) showed from Nd isotope systematics that early (pre 2.71 Ga) magmas of the Oxford Lake volcanic assemblage did not interact with older crust and hence, must have developed in juvenile oceanic or continental margin arcs. Hence, the depleted basalts at Ponask Lake and the Muskrat Dam assemblage appear to represent obducted slices of juvenile oceanic material and continental margin arc. These provide an indication that oceanic basins separated older crustal fragments within what is now the Island Lake terrane until after 2.73 Ga.

In summary, the Island Lake terrane is characterized by magmatic rocks, whose ages span most of Geon 28. At opposite ends of the age spectrum, these include mafic volcanic sequences of the Whiteway assemblage at Island Lake with an interpreted age of 2.90 Ga (Parks et al. 2003) and nearby gabbro with an age of 2.81 Ga (No. 13 of Table 1). The majority of volcanic rocks have a narrower age range from 2.86 to 2.85 Ga represented by the Pipe Point assemblage at Island Lake (Parks et al. 2003) and the Sachigo assemblage at Ponask Lake (No. 8 of Table 3). Many volcanic sequences and plutonic complexes of the Island Lake terrane have been emplaced through older crust of ca 3.0 Ga age. Although the older crust may represent the North Caribou terrane, greenstone sequences apparently derived from juvenile oceanic sources and continental margin arcs are interspersed with the contaminated rocks and suggest that the Island Lake terrane is a collage of highly reworked crustal blocks.

Possibly major shear zones such as the Savage Island shear zone at Island Lake and a major fault at the north side of the North Caribou greenstone belt (F.W. Breaks, Geoscientist, Ontario Geological Survey, personal communication, 2004) mark the south margin of the Island Lake terrane. Parks et al. (2003) provided evidence that the Savage Island shear zone was active at 2.70 Ga, an age that may represent docking of the Island Lake terrane with the North Caribou terrane or late ensialic faulting. In the former case, volcanic sequences south of the faults such as the Jubilee assemblage and North Rim assemblage could have developed autochthonously on the North Caribou terrane. The north margin of the Island Lake terrane is interpreted to extend from south of Cross Lake through southern Knee Lake and Gods Lake Narrows to southern Stull Lake and Big Trout Lake (*see* Figure 3). In detail, the north margin of the Island Lake terrane follows the Stull Lake–Wunnummin Lake fault in the area of Stull Lake and mafic sequences of the Stull–Swan lakes greenstone belt occurring south of the fault are part of the Island Lake terrane (*see* map in back pocket). The Island Lake terrane is larger than was envisaged by Thurston, Osmani and Stone (1991) and includes much of the former Munro Lake and Muskrat Dam terranes by virtue of Geon 28 plutonic rocks in these areas.

OXFORD LAKE–STULL LAKE TERRANE

Volcanic sequences of the Hayes River assemblage are dated at 2.83 Ga at Knee Lake (Nos. 5 to 7 of Table 1) and represent the oldest known rocks of the Oxford Lake–Stull Lake terrane. Sedimentary sequences are associated with the Hayes River assemblage and include greywacke and the Opischikona sediments at Knee Lake. The sedimentary sequences contain detrital zircon grains ranging in age from 2.82 to 2.97 Ga (Nos. 8 and 9 of Table 1) indicating that the rocks were deposited at least a few Ma after eruption of Hayes River lavas but also derived material from older sources.

Despite evidence for pre 2.83 Ga crustal sources for sedimentary sequences associated with the Hayes River assemblage, recent geochronologic investigations have failed to identify old plutonic rocks in the Oxford Lake–Stull Lake terrane. Plutonic rocks of the present area range in age from 2.73 to 2.71 Ga and only one sample shows evidence of having inherited significantly older material of 2.84 Ga age (Nos. 20, 21, 22 and 24 of Table 3). The isotopic studies of Skulski et al. (2000) show that plutonic rocks of the Oxford Lake–Stull Lake terrane in the present area have positive ε_{ND} values and model ages of 2.71 to 2.78 Ga indicating that they have not interacted with older crust as occurred in the Island Lake terrane. These data are interpreted to indicate that the Hayes River assemblage and other assemblages representative of ca 2.83 Ga mafic sequences comprise exotic material that was tectonically transported into position prior to 2.73 Ga. Enriched and depleted varieties of basalts (e.g., Syme et al. 1999) and pre 2.83 Ga detrital zircon grains in associated sedimentary units suggest that the mafic sequences developed in magmatic arc or ocean floor environments on or near the margin of the Island Lake terrane.

Early, mafic to felsic calc-alkalic sequences of the Oxford Lake volcanic assemblage erupted at 2.73 to 2.72 Ga in the Oxford Lake–Stull Lake terrane (Nos. 17 to 21 of Table 1 and Nos. 10 to 14, 19, 25 and 26 of Table 3). Magma compositions became more alkalic until after 2.71 Ga, when volcanism ceased and the volcanic sequences were unconformably overlain by the Oxford Lake sedimentary assemblage and Cross Lake assemblage. Here again, the Nd isotope studies of Skulski et al. (2000) have been instrumental in modeling tectonic evolution by showing that the early calc-alkalic phases of the Oxford Lake volcanic assemblage are isotopically juvenile whereas the late alkalic phases are evolved. The calc-alkalic volcanism evidently occurred in oceanic or continental margin arcs with minimum interaction of older crust whereas the alkalic magmas passed through and assimilated older material possibly representative of the Northern Superior Superterrane.

Late sedimentary sequences of the Oxford Lake sedimentary assemblage and Cross Lake assemblage were deposited in marine basin and alluvial-fluvial environments during and after eruption of the Oxford Lake volcanic assemblage. The youngest detrital zircon grains indicate that deposition occurred after 2706 Ma at Oxford Lake, 2713 Ma at Little Stull Lake and 2707 Ma at Knee Lake (Nos. 22, 23, 26, 27 of Table 1). In view of recent changes in assemblage names, it is often unclear however whether the geochronology applies to what is now known as the Oxford Lake sedimentary assemblage or the Cross Lake assemblage. In the present area, the youngest zircon grains in two samples of wacke representative of the Oxford Lake sedimentary assemblage are 2.72 Ga whereas those in arkosic samples representative of the Cross Lake assemblage are 2.71 Ga. As discussed further below in relation to the geology of the Stull–Swan lakes greenstone belt, these data indicate that deposition of the Oxford Lake sedimentary assemblage could have occurred earlier than the Cross Lake assemblage. Volcanism and sedimentation appears to have evolved from an oceanic or continental margin arc environment at 2.73 to 2.72 Ga to a continental environment at 2.71 Ga, possibly in response to accretion of the Northern Superior Superterrane to the north. Rare Paleoarchean zircon grains in late sedimentary sequences at Cross Lake and Knee Lake (Nos. 16 and 27 of Table 1) suggest that the Northern Superior Superterrane shed detritus into sedimentary sequences of the Oxford Lake–Stull Lake terrane after 2713 Ma.

In summary, the Oxford Lake–Stull Lake terrane represents a mature volcanic arc that developed on or near the margin of the Island Lake terrane at 2.73 Ga. Old mafic sequences such as the Hayes River assemblage were thrust onto the continental margin or onto early, mature parts of island arcs represented by 2.73 Ga biotite tonalite batholiths (No. 24 of Table 3) at an initial stage in evolution of the Oxford Lake–Stull Lake terrane. Volcanism with associated sedimentation in marine basins and extensive plutonism continued until 2.71 Ga at which time docking of the Northern Superior Superterrane caused overall shortening of the continental margin and activated regional strike-slip faults. At or shortly after 2.71 Ga, alkalic volcanism and coarse clastic sedimentation took place, commonly within basins distributed along the strike-slip faults.

The south boundary of the Oxford Lake–Stull Lake terrane extends from south of Cross Lake to south of Oxford Lake through Gods Lake Narrows and hence eastward along the Stull Lake–Wunnummin Lake fault through Stull Lake to Big Trout Lake (*see* Figure 3 and map in back pocket). In the present area, the north boundary is marked by the South Kenyon fault. Extreme western and eastern extents of this terrane, particularly east of Big Trout Lake, are poorly known.

NORTHERN SUPERIOR SUPERTERRANE

The Northern Superior Superterrane represents highly assimilated Paleoarchean crust at the north margin of the Superior Province. Although evidence of ancient crust came initially from northern Manitoba (e.g., Krogh et al. 1986), geochronologic and isotopic studies from the present area and the adjacent Red Cross Lake area of Manitoba have greatly added to the information available on the Northern Superior Superterrane. In this area, plutonic rocks of tonalite to granodiorite composition show crystallization ages of 2.85, 2.82, 2.81 and 2.78 Ga (Nos. 36 to 38 of Table 3 and data of Corkery and Stone, in press). Two samples of tonalite gneiss have inherited zircon grains with cores ranging in age from 3.57 to 3.21 Ga. Volcanic rocks identical in age to the 2.83 Ga Hayes River assemblage at Knee Lake have been identified at Dadson Lake and Red Cross Lake (No. 30 of Table 3 and data of Corkery and Stone, in press). These volcanic sequences occur along the north side of the North Kenyon fault (see map in back pocket) and their structural relation to the Northern Superior Superterrane is unclear. The volcanic sequences could have been faulted into position although the Dadson Lake sample contains 2.85 Ga inherited zircon grains, suggesting that the volcanic magma may have erupted through and interacted with older basement material.

The Northern Superior Superterrane was affected by a late magmatic event from 2.72 to 2.69 Ga, when voluminous plutons and dikes of tonalite to granodiorite were emplaced (Nos. 33, 34, 35, 39 and 41 of Table 3). Small units of felsic tuff and associated wacke of 2.72 Ga age (Nos. 31 and 42 of Table 3) occur along the north side of the North Kenyon fault. Also spatially associated with the major fault and located 20 km east of Yelling Lake (see map in back pocket) is a small gabbro body and chemically similar mafic volcanic rocks of 2.716 Ga age. The trace element characteristics of the mafic rocks are comparable to oceanic basalts, which invites the interpretation that the mafic sequences developed in an oceanic environment and were obducted onto the south margin of the Northern Superior Superterrane after 2.716 Ga. Similarly, the 2.72 Ga tuff and wacke sequences, which can be correlated with the Oxford Lake volcanic and sedimentary assemblages, may have been tectonically transported onto the Northern Superior Superterrane.

Skulski et al. (2000) investigated the Nd isotopic composition of felsic plutonic rocks including those that crystallized from 2.85 to 2.81 Ga and from 2.72 to 2.71 Ga in the Northern Superior Superterrane. All show negative ε_{ND} values and Nd model ages ranging from 2.9 to 3.6 Ga. These provide convincing evidence that the plutonic suites of the Northern Superior Superterrane have been derived by partial melting of older crust, some of which is Paleoarchean in age. Skulski et al. (2000) also noted growth rims on zircon grains and dated these using a sensitive high-resolution ion microprobe (SHRIMP). Zircon grains from older plutonic rocks (2.85 to 2.81 Ga) had overgrowths dated at 2.74 and 2.76 to 2.72 Ga, whereas zircon grains from the youngest pluton (2.71 Ga) had no overgrowths. The overgrowths are interpreted as having developed during one or more metamorphic events that affected the Northern Superior Superterrane between 2.74 and 2.71 Ga.

In Manitoba, the Northern Superior Superterrane includes the Split Lake block, Pikwitonei granulite domain and possibly the Orr Lake block, but does not include the exotic Paleoarchean Assean Lake block (*see* Figure 3). The evidence for ancient crust is scarce in the Split Lake, Pikwitonei and Orr Lake blocks due to pervasive Neoarchean magmatism, deformation and metamorphism.

The Split Lake block is underlain by metamorphosed tonalite to granodiorite with lesser components of granite and gabbro. The geochronologic study of Böhm et al. (1999) identified two magmatic events at 2.84 Ga and 2.71 Ga. The core of a zircon grain within a sample of granodiorite was found to be older than 3.0 Ga. The Split Lake block was affected by three periods of metamorphism at 2.705, 2.695 and 2.62 Ga as shown by overgrowths on zircon grains. The Pikwitonei granulite domain is dominated by orthopyroxene-bearing tonalite and granodiorite containing remnants of supracrustal rocks. Krogh et al. (1986), Heaman, Machado and Krogh (1986) and Böhm (1998) reported pre 2.9 Ga zircon cores from tonalite gneisses although most rocks have emplacement ages that cluster around 2.7 Ga. At least two distinct metamorphic zircon growth events diachronously affected various parts of the Pikwitonei domain from 2.695 to 2.64 Ga. The Orr Lake Block is highly deformed and less well studied. Böhm et al. (1999) suggest that it may represent a mix of Archean and Proterozoic rocks.

The crystallization ages of plutonic rocks and the ages of inherited zircon grain from the Split Lake block comparable favourably to those of the Northern Superior Superterrane in the present area and demonstrate that both areas have been affected by Mesoarchean and Neoarchean magmatic events. Further, both areas show evidence of older inherited material. The Pikwitonei domain appears to lack ca

2.84 Ga magmatism or else evidence for the event has not yet been identified. In either case, Heaman, Böhm and Creaser (1998) and Böhm (1998) showed 3.1 to 3.5 Ga Nd model ages for samples from the Split Lake as well as the Pikwitonei blocks. This strongly suggests that the Split Lake and Pikwitonei blocks of northern Manitoba form the western extension of the Northern Superior Superterrane.

Scattered evidence of older crust, possibly correlative with the Northern Superior Superterrane has been found in the Superior Province of northern Québec. In the north-trending Lake Minto and Goudalie blocks of northern Québec, tonalite gneiss enclaves with crystallization ages of 3.15 and 3.01 Ga have been found (Percival et al. 1992), the older of which contains 3.5 Ga inherited zircon grains. The gneissic enclaves and some 2.7 Ga diatectic plutonic suites in the area have negative ε_{ND} values, indicating their magmas interacted with older crustal material (Stern et al. 1994). Simard et al. (2002) reported an age of 3.83 Ga for an enclave of volcanic rocks in the Inukjuak block on the eastern side of Hudson Bay.

Although correlation between the northwestern and northeastern Superior Province is rendered difficult by lack of exposure, Percival et al. (1994) suggested a broadly arcuate distribution of terranes from west to east. In this model, the 3.0 Ga North Caribou terrane is extended east through James Bay and hence north to connect with assimilated Mesoarchean crust in the Lake Minto and Goudalie domains. Similarly, the Northern Superior Superterrane extends east through southern Hudson Bay and hence, north to possibly correlate with Paleoarchean crust on the eastern side of Hudson Bay.

In summary, the Northern Superior Superterrane is made up of several generations of felsic plutonic and gneissic rocks with rare greenstone belts and supracrustal enclaves. Felsic intrusions compositionally variable from tonalite through granodiorite to granite and mafic to felsic volcanic sequences developed within one or more magmatic events that occurred broadly in the interval of 2.85 to 2.78 Ga and from 2.72 to 2.69 Ga. Although no Paleoarchean rocks have yet been identified, inherited zircon grains and Nd isotope characteristics suggest that the felsic intrusive magmas were partially melted from or interacted with older crust of 3.57 to 3.21 Ga age. Some Neoarchean supracrustal sequences appear to have developed in an oceanic environment and were tectonically emplaced at the south margin of the Northern Superior Superterrane. Several Neoarchean metamorphic events are recorded by growth rims on zircon grains and appear to have diachronously affected the area. Metamorphic grade of all but the youngest plutonic rocks ranges from amphibolite to granulite facies.

In the west, the Northern Superior Superterrane includes the Pikwitonei and Split Lake blocks that are in fault contact with the Paleoarchean Assean Lake block and mixed Proterozoic and Archean rocks of the Trans Hudson Orogen. Northward, the Northern Superior Superterrane is overlain by the Paleoproterozoic Fox River sill (Scoates 1990) and Paleozoic sequences of the Hudson Bay Lowland. In the present area, the south boundary of the Northern Superior Superterrane is marked by the South Kenyon fault (see map in back pocket). The eastern extension of this terrane boundary fault is poorly known.

TERRANE INTERACTION

The Island Lake, Oxford Lake–Stull Lake and Northern Superior terranes evolved independently through various parts of Paleoarchean, Mesoarchean and Neoarchean time and were tectonically joined about 2.7 billion years ago. A combination of geochronology, Nd isotope systematics and Al-in-hornblende barometry constrains the timing of terrane amalgamation as well as relative vertical motion of the crustal blocks as they came together.

Depleted basalts of the 2.716 Ga Yelling assemblage are interpreted to represent remnants of an ocean floor at the south side of the Northern Superior Superterrane. The lack of chemical evidence for interaction between the mafic magmas and older crust as well as the fault-bounded nature of the Yelling assemblage suggests that the oceanic material was tectonically transported onto the south margin of the Northern Superior Superterrane after 2.716 Ga.

Al-in-hornblende barometry (*see* "Aluminum-in-Hornblende Barometry") provides insight on the relative vertical motions of the Northern Superior and Oxford Lake–Stull Lake blocks from 2.716 to 2.714 Ga. A pluton of the hornblende tonalite suite extends from 25 to 50 km east of Yelling Lake in the Northern Superior Superterrane (*see* map in back pocket) and is dated at 2.716 Ma (No. 35 of Table 3). Three Al-in-hornblende pressure determinations ranging from 2.1 to 3.2 kbars (average of 2.7 kbars) constrain the depth of crystallization of this pluton[3]. In contrast, a pluton of the sanukitoid suite centered 20 km east of Yelling Lake is dated at 2.714 Ga (No. 33 of Table 3) and has four Al-in-hornblende pressure determinations ranging from 2.9 to 3.7 kbars (average of 3.3 kbars). These data are interpreted to indicate that the south margin of the Northern Superior Superterrane moved down slightly from 2.716 to 2.714 Ga. The vertical displacement is on the order of 1 to 2 km corresponding to the 0.6 kbar difference in pressure between average pressures for the hornblende tonalite pluton and the sanukitoid pluton.

The Oxford Lake–Stull Lake terrane can be subdivided into two blocks on the basis of Al-in-hornblende barometry. The northern block (Block 3) is somewhat tapered in plan and extends from the South Kenyon fault to the Ellard Lake greenstone belt. The southern block (Block 2) extends south from the Ellard Lake greenstone belt to the Stull–Wunnummin fault (*see* map in back pocket for locations of these features). In the northern block, two plutons of the hornblende tonalite suite have ages of 2.715 and 2.719 Ga (Nos. 20 and 22 of Table 3). These and related hornblende tonalite intrusions have Al-in-hornblende pressure determinations ranging from 2.9 to 5.8 kbars (average of 4.1 kbars). Three plutons of the sanukitoid suite, one of which is dated at 2.71 Ga (No. 21 of Table 3), have Al-in-hornblende pressure determinations of 1.1[4] to 2.0 kbars (average of 1.3 kbars). The difference in average pressure between plutons of the hornblende tonalite and sanukitoid suites (4.1-1.3 kbars) represents an uplift of approximately 8 km in the northern part of the Oxford Lake–Stull Lake terrane from 2.719 to 2.71 Ga.

The southern block of the Oxford Lake–Stull Lake terrane is characterized by intrusions of the hornblende tonalite suite showing Al-in-hornblende pressures in the range of 6.5 to 9.6 kbars (average of 8.5 kbars). The pressure determinations for hornblende tonalite are considerably higher in the southern block than in the northern block and suggest a south-side-up displacement on the order of 13 km. This vertical displacement of the south block relative to the north block evidently occurred along a discontinuity in the vicinity of the Ellard Lake greenstone belt.

Skulski et al. (2000) constrained the timing of interaction between the Oxford Lake–Stull Lake terrane and the Island Lake terrane (the latter of which they referred to as the Munro Lake terrane) using geochronology and Nd isotopes. These authors noted that the 2.71 Ga Rorke Lake pluton of the sanukitoid suite has a juvenile isotopic character indicating that the magma did not interact with older crustal material. In contrast, alkalic volcanic rocks in the Stull Lake area with an age inferred to be younger than 2.71 Ga have an evolved isotopic ratio. Further, the Cross Lake assemblage was deposited after 2.71 Ga and contains a wide range of clast lithologies and detrital zircon grains representative in age of the Island Lake and Northern Superior terranes as well as the Oxford Lake–Stull Lake terrane. These data were interpreted to indicate that the Oxford Lake–Stull Lake terrane was joined to the Island Lake terrane probably within a few million years after 2.71 Ga.

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[3] See the pluton outline and pressure data in Figure 19a.

[4] The pressure is determined using the method of Schmidt (1992).

Late alkalic magmas that erupted after 2.71 Ga along the terrane boundary zone would have passed through and assimilated older crust of the Island Lake terrane accounting for the evolved Nd isotopic ratio of this material. Sedimentary detritus deposited after 2.71 Ga could have been derived from any of the amalgamated terranes, thus explaining the wide range of zircon ages in the Cross Lake assemblage (e.g., Nos. 16 and 27 of Table 1).

Al-in-hornblende pressure determinations on intrusions of the hornblende tonalite suite range from 1.0 to 5.8 kbars with an average of 4.6 kbars within Block 1 extending south from the Stull–Wunnummin fault to the Ponask Lake greenstone belt. With the exception of one extreme value of 1.0 kbars, pressure estimates do not vary systematically south of the Stull–Wunnummin fault, including the area south of the Ponask Lake greenstone belt, which indicates that the margin of the Island Lake terrane did not bend or tilt during terrane interaction. Rather, the southern block of the Oxford Lake–Stull Lake terrane was ramped upward through a vertical distance of 8 km as indicated by the difference in average crystallization pressures for hornblende-bearing plutonic rocks between the southern Oxford Lake–Stull Lake terrane and the Island Lake terrane (8.5-4.6 kbars).

Terrane assembly seems to have proceeded from north to south. Although an ocean may have separated the Northern Superior Superterrane from the Oxford Lake–Stull Lake terrane at 2.716 Ga, the margins of these terranes were being down-thrown and uplifted by 2.714 Ga. The vertical displacements of terrane margins are best explained by abutment and thrusting of one block onto the other, in this case the Oxford Lake–Stull Lake terrane was ramped up onto the Northern Superior Superterrane (Figure 4). Hence, terrane abutment had at least initiated in the north by 2.714 Ga. Southward, it has been demonstrated that the Oxford Lake–Stull Lake terrane abutted and overthrusts the Island Lake terrane shortly after 2.71 Ga. Notably, the Oxford Lake–Stull Lake terrane was underthrust and uplifted at both margins and was also shortened by displacement on an internal fault along the Ellard Lake greenstone belt (*see* Figure 4).

Major faults such as the North Kenyon, South Kenyon and Stull–Wunnummin faults became active after terrane assembly and are commonly localized along boundaries of amalgamated terranes. Horizontal mineral lineations and a variety of kinematic indicators imply predominantly dextral transcurrent displacement on these structures. Late sedimentary sequences such as the Cross Lake assemblage show a strong spatial association with major faults and probably were deposited in pull-apart basins localized along the transcurrent faults. Parks et al. (2003) provided geochronologic evidence that a major fault at Island Lake was active as late as 2.70 Ga. This age may mark the final stages of terrane accretion and transpressive deformation.

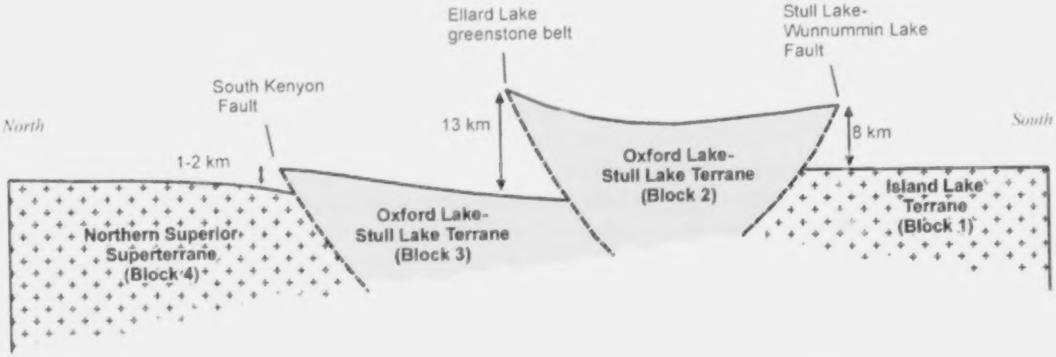


Figure 4. Schematic vertical section through the Northern Superior, Oxford Lake–Stull Lake and Island Lake terranes showing vertical displacement on boundary faults.

Data: Rock Samples, Rock Compositions and Mineral Compositions

Approximately 660 rock samples were collected from the northern Superior area to aid in the geologic interpretation of the area. Of these, 197 samples were analyzed for major elements, including 188 samples that were also analyzed for a range of trace elements. In addition, many samples were assayed for selected metals of economic value. Polished thin sections were made from samples representative of the major rock types for petrographic analysis and determination of mineral compositions by microprobe. Amphiboles and feldspars were analyzed extensively for Al-in-hornblende barometry and thermometry (discussed in "Aluminum-in-Hornblende Barometry").

Digital data files of the analytical results, listed in Table 4, are contained on Miscellaneous Release–Data 135 (MRD 135), available separately from this report. Representative pieces of the rock samples, powders of the samples used for chemical analysis and the polished thin sections are stored in the archives of the Ontario Geological Survey in Sudbury.

Table 4. List of data files on Miscellaneous Release–Data 135 (MRD 135), available separately from this report. Files of the rock sample list, rock compositions and mineral compositions are in Microsoft® Excel (.xls) and comma-delimited (.csv) format; the lithologic legend is in Microsoft® Word (.doc) and PDF (Adobe® Acrobat®) format.

File Name	File Size (kilobytes)	Contents
Amphibole.xls / Amphibole.csv	176 / 46	90 amphibole analyses with Al-in-hornblende pressure determinations
Biotite.xls / Biotite.csv	24 / 4	13 biotite analyses
Chlorite.xls / Chlorite.csv	23 / 3	8 chlorite analyses
Feldspar.xls / Feldspar.csv	81 / 29	134 feldspar analyses
Garnet.xls / Garnet.csv	35 / 9	25 garnet analyses
Lithologic legend.doc / Lithologic legend.pdf	34 / 70	Detailed lithologic legend with explanation of rock type codes used on maps and data files
Muscovite.xls / Muscovite.csv	20 / 2	3 muscovite analyses
North Superior Geochemistry 2000.xls / North Superior Geochemistry 2000.csv	226 / 75	Major elements and trace elements of 182 rock samples
Pyroxene.xls / Pyroxene.csv	26 / 4	13 pyroxene analyses
Spinel.xls / Spinel.csv	22 / 3	6 spinel analyses
Staurolite.xls / Staurolite.csv	17 / 1	4 staurolite analyses
Samples North Superior 2004.xls / Samples North Superior 2004.csv	155 / 50	List of approximately 660 rock samples showing locations, rock type, assay results, stain results, mineral assemblage and other uses

Geology and Geochemistry

GENERAL GEOLOGY

The northern Superior area is underlain by four east-southeast-trending Archean greenstone belts interspersed with broad felsic plutonic domains. From south to north, the greenstone belts include the Ponask Lake greenstone belt, Stull–Swan lakes greenstone belt, Ellard Lake greenstone belt and Yelling

Lake greenstone belt (*see* Figure 2). Supracrustal strata of the greenstone belts are divided into assemblages, which are discussed further below.

Pillowed to massive mafic lavas with minor components of gabbro, komatiite and intermediate to felsic volcanic fragmental rocks are characteristic of the Pierce and Sachigo assemblages in the Ponask Lake greenstone belt. These and associated thin units of wacke, tonalite-cobble conglomerate and marble were deposited at about 2.85 Ga (*see* Table 3). The Stull–Swan lakes greenstone belt is dominated by several chemically distinct assemblages of mafic volcanic rocks with minor gabbro. These include the Rorke Lake, Edmund Lake, Stull and Swan assemblages. The undated mafic sequences are overlain by mafic to intermediate and felsic flows and fragmental rocks of the Oxford Lake volcanic assemblage. Eruption of the Oxford Lake volcanic assemblage occurred from 2.73 to 2.71 Ga. Deposition of thin-bedded wacke and coarse sandstone and conglomerate sequences of the Oxford Lake sedimentary and the Cross Lake assemblages occurred during and after the late volcanism. The sedimentary sequences occupy the central part of the Stull–Swan lakes greenstone belt at Stull Lake and occur as thin units in the area of Swan Lake (see map in back pocket).

Mafic volcanic sequences of the Rorke Lake and Ellard assemblages comprise the major part of the Ellard Lake greenstone belt. Intermediate to felsic volcanic sequences of the Oxford Lake volcanic assemblage occur along the central axis of the Ellard Lake greenstone belt at Ellard Lake and are dated at 2.73 Ga. A unit of coarse conglomerate with associated wacke of the Cross Lake assemblage occurs west of Ellard Lake. The Yelling Lake greenstone belt is composed of several bifurcated and metamorphosed slivers of mafic volcanic rocks. These include the 2.83 Ga Dadson assemblage and the 2.72 Ga Yelling assemblage as well as an assemblage of unknown age. Thin units of felsic tuffs and wacke of the Oxford Lake volcanic and sedimentary assemblages occur locally.

Felsic plutonic rocks occur widely in the area and are subdivided into six suites (see Table 5 on p.46) of which the biotite tonalite, biotite granite and hornblende tonalite suites are the most voluminous. These occur as large elongate batholiths between greenstone belts. Quartz-undersaturated rocks of the sanukitoid suite are compositionally variable from monzodiorite through monzonite to granite and occur as oval plutons typically intruding greenstone belts. Thin belts of gneissic tonalite occur sporadically and rare units of two-mica granite of the peraluminous suite are associated with sedimentary rocks in the Ponask Lake and Stull–Swan lakes greenstone belts.

The northern Superior area is cut by two major east-southeast-trending faults. The Stull–Wunnummin fault occurs at the south side of the Stull–Swan lakes greenstone belt and the northern and southern splays of the Kenyon fault occur south of the Yelling Lake greenstone belt (*see* Figure 5). The faults are marked by broad mylonite zones and are interpreted to demarcate terrane boundaries. The Island Lake terrane lies south of the Stull–Wunnummin fault whereas the Oxford Lake–Stull Lake terrane extends north from the Stull–Wunnummin fault to the Kenyon fault and the Northern Superior Superterrane is situated north of the Kenyon fault.

Two sets of gabbro dikes trend north-northeast and northwest through the northern Superior area. The sets of dikes are correlated with the Proterozoic Molson Lake Swarm and MacKenzie Swarm, respectively (Ontario Geological Survey 1991). Proterozoic magmatism is also represented by an oval carbonatite intrusion north of McLeod Lake (*see* Figure 5). Archean and Proterozoic rocks of the northern Superior area are onlapped to the north by Ordovician limestone of the Hudson Bay Lowland.

Descriptions of the geology and geochemistry of the greenstone belts, felsic plutonic suites and dike swarms are provided below.

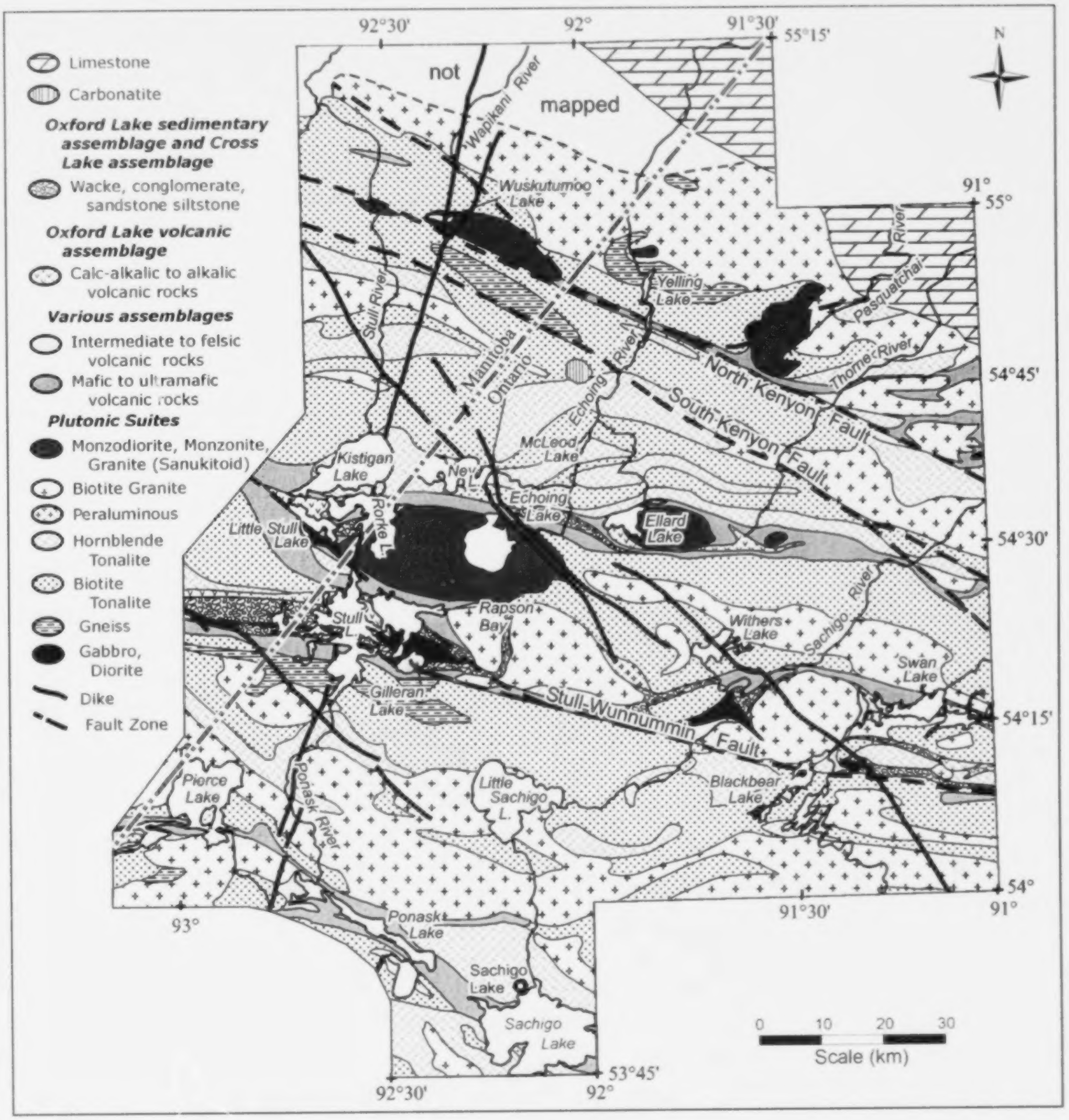

Limestone
Carbonatite
Oxford Lake sedimentary assemblage and Cross Lake assemblage
Wacke, conglomerate, sandstone siltstone
Oxford Lake volcanic assemblage
Calc-alkalic to alkalic volcanic rocks
Various assemblages
Intermediate to felsic volcanic rocks
Mafic to ultramafic volcanic rocks
Plutonic Suites
Monzodiorite, Monzonite, Granite (Sanukitoid)
Biotite Granite
Peraluminous
Hornblende Tonalite
Biotite Tonalite
Gneiss
Gabbro, Diorite
Dike
Fault Zone
92°30'
92°
91°30'
55°15'
N
not
mapped
Wapikani River
Wuskutumoo Lake
91°
55°
Yelling Lake
Pasquatchai River
Stull River
Manitoba
Ontario
Echoing River
North Kenyon Fault
South Kenyon Fault
Thorne River
54°45'
McLeod Lake
Kistigan Lake
New L.
Echoing Lake
Ellard Lake
Little Stull Lake
Rorke L.
54°30'
Rapson Bay
Stull L.
Withers Lake
Sachigo River
Swan Lake
Gilleran Lake
Stull-Wunnummin Fault
54°15'
Pierce Lake
Ponask River
Little Sachigo L.
Blackbear Lake
54°
93°
91°30'
91°
Ponask Lake
Sachigo Lake
Sachigo Lake
0
10
20
30
Scale (km)
53°45'
92°30'
92°

Figure 5. Geology of the northern Superior area.

PONASK LAKE GREENSTONE BELT

The Ponask Lake greenstone belt extends discontinuously east-southeast from Red Sucker Lake, Manitoba through Pierce, Ponask and Sachigo lakes, Ontario over a distance of 150 km and underlies approximately 2.7% of the northern Superior area. Width of the greenstone belt ranges from several hundred metres to a maximum of 10 km at Sachigo Lake and is generally in the range of 1 to 3 km. The Ponask Lake greenstone belt can be divided into two belts separated by several hundred metres of tonalite at northwest Ponask Lake. The western segment (Pierce assemblage) tapers to a narrow arm extending *en echelon* to the north of the Sachigo assemblage at northwest Ponask Lake. Narrow parts of the greenstone belt, particularly the Pierce assemblage, tend to be more highly strained than wide parts of the belt and are composed of amphibole and biotite schists. In contrast, basaltic flows with large undeformed pillows are observed on islands in Sachigo Lake where the belt is widest.

Pierce Assemblage

The Pierce assemblage is composed of mafic volcanic sequences with lesser units of wacke and marble. All strata are strongly foliated and the sedimentary rocks are locally migmatized and intermingled with coarse white two-mica granite. Quartz-rich wacke at Pierce Lake has a single population of zircon grains dated at 2863 Ma (No. 6 of Table 3). This date is interpreted to represent the maximum age of the sedimentary and associated volcanic sequences. A minimum age of 2723 Ma for the Pierce assemblage is provided by the intrusive biotite granite pluton south of Pierce Lake (No. 2 of Table 3).

The composition of the mafic volcanic sequences is mainly basalt and basaltic komatiite with rare komatiite (Figure 6a). All volcanic rocks are subalkalic and belong to the low- to medium-K (tholeiite and calc-alkaline series) of Figure 7a. The majority of basalt and basaltic komatiite samples are enriched from 2 to 30 times primitive mantle values in trace elements (Figure 8a) and typically show Th>La>Nb. One sample is classified as depleted basalt and has Th≈Nb≈La (open circles of Figure 8b). Two samples of the enriched mafic volcanic rocks (97DS43 and 97DS51) have 55% SiO_2 and are classified as siliceous high-Mg basalts (Arndt and Jenner 1986).

Sachigo Assemblage

The Sachigo assemblage is somewhat more lithologically diverse than the Pierce assemblage. It contains a spectrum of mafic volcanic rocks ranging compositionally from komatiite through basaltic komatiite to basalt (*see* Figure 6a) and dacite and rhyolite. The felsic volcanic rocks occur as thin units of tuff and volcanic breccias and quartz+feldspar porphyritic intrusive rocks within mafic volcanic strata at Ponask Lake and broaden within a large oval domain at western Ponask Lake. The mafic volcanic rocks belong mainly to the low-K series although felsic volcanic rocks scatter into the medium-K field of Figure 7a.

At northwestern Ponask Lake, biotite tonalite is overlain by conglomerate with tonalite clasts and sequences of komatiite intermixed with massive to layered magnesite+serpentine+talc, the latter of which possibly represents assimilated marble sequences. The conglomerate and komatiite are also mapped at scattered localities along the north side of the Sachigo assemblage at central Ponask Lake and Sachigo Lake.

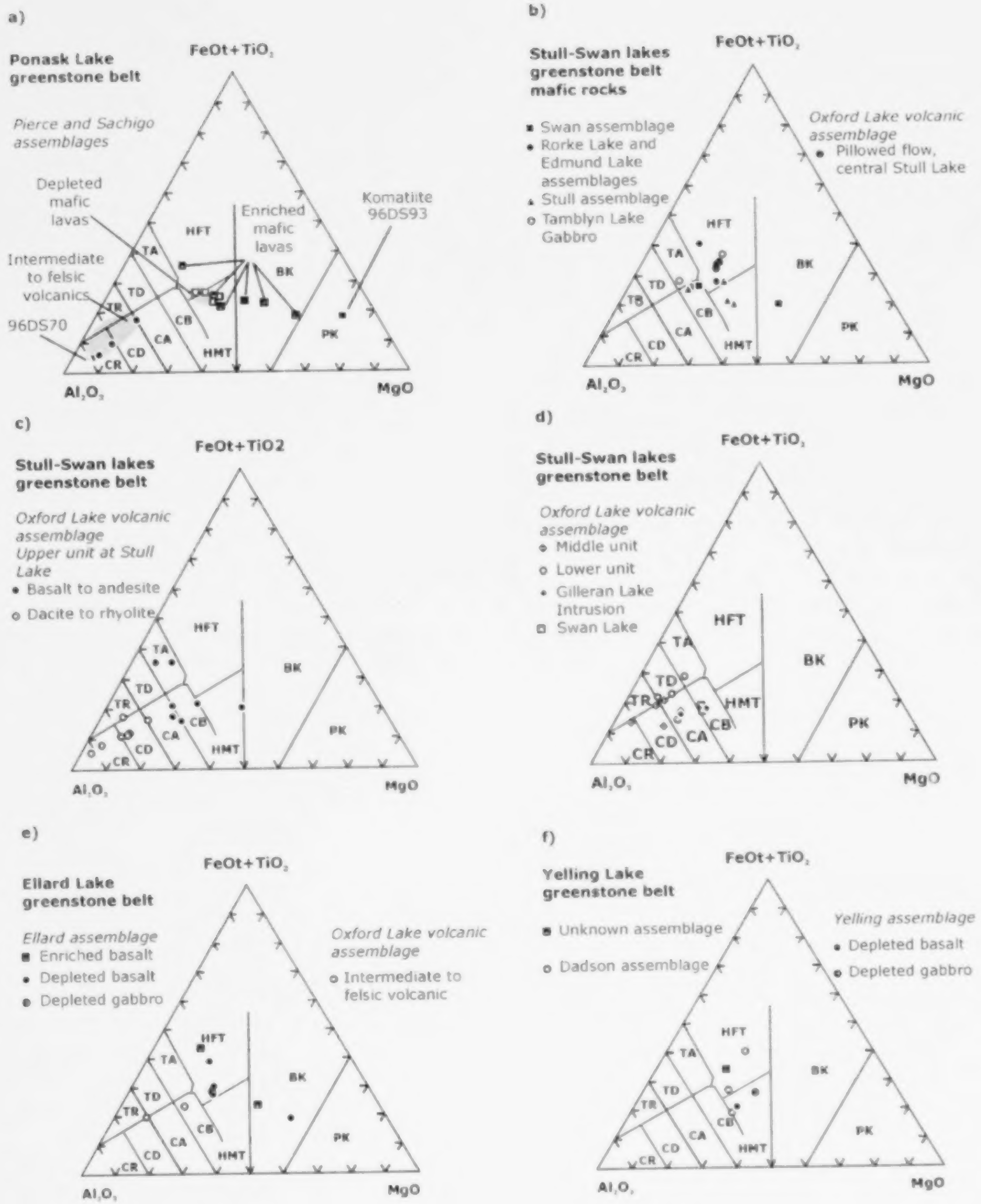


Figure 6. Cation plot of Jensen (1976) for volcanic rocks of all greenstone belts. PK - komatiite; BK - basaltic komatiite; HMT - high-magnesium tholeiite; HFT - high-iron tholeiite; CB - calc-alkalic basalt; TA - tholeiitic andesite; CA - calc-alkalic andesite; TD - tholeiitic dacite; CD - calc-alkalic dacite; TR - tholeiitic rhyolite; and, CR - calc-alkalic rhyolite.

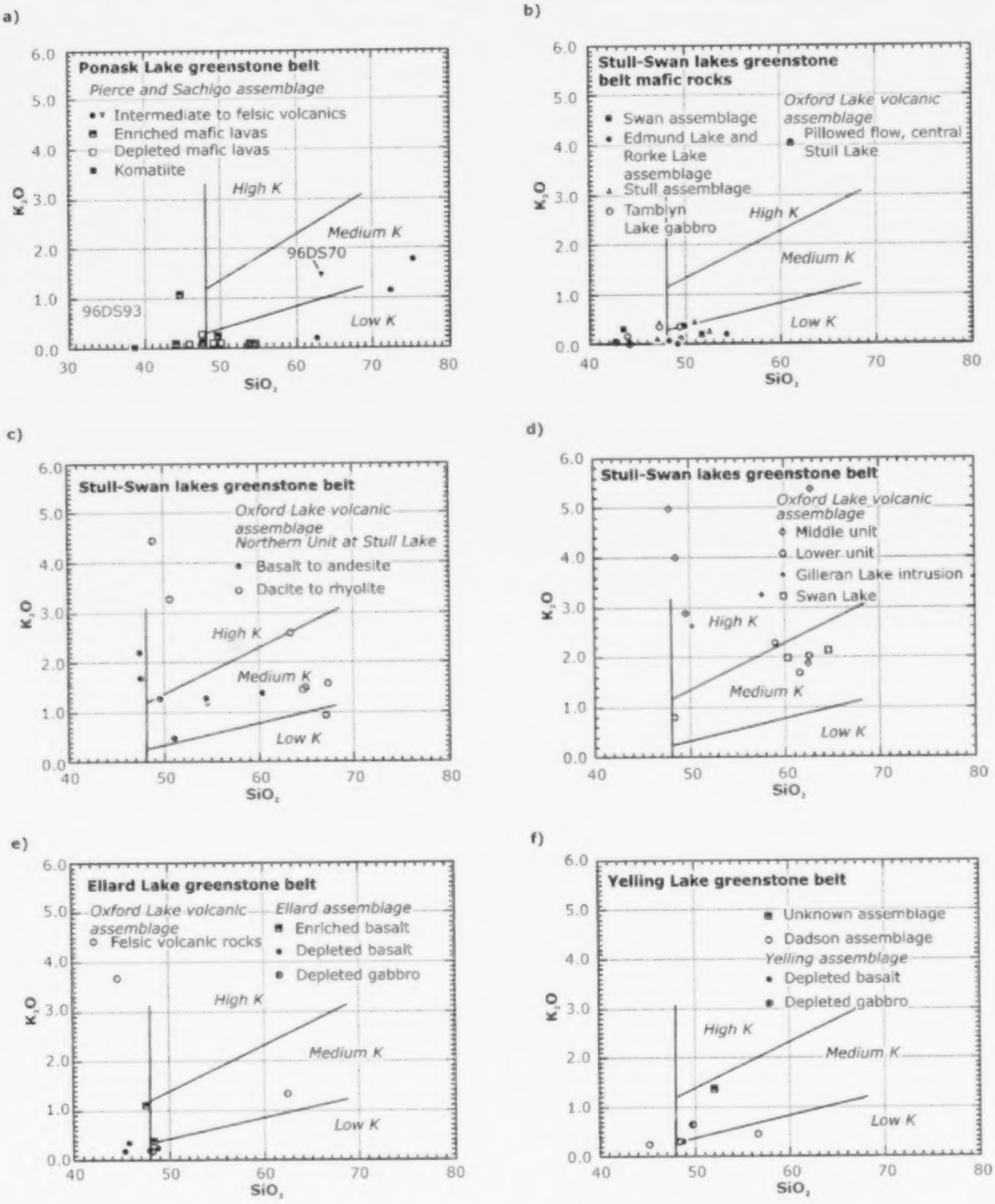


Figure 7. K_2O-SiO_2 plots for volcanic rocks of all greenstone belts.

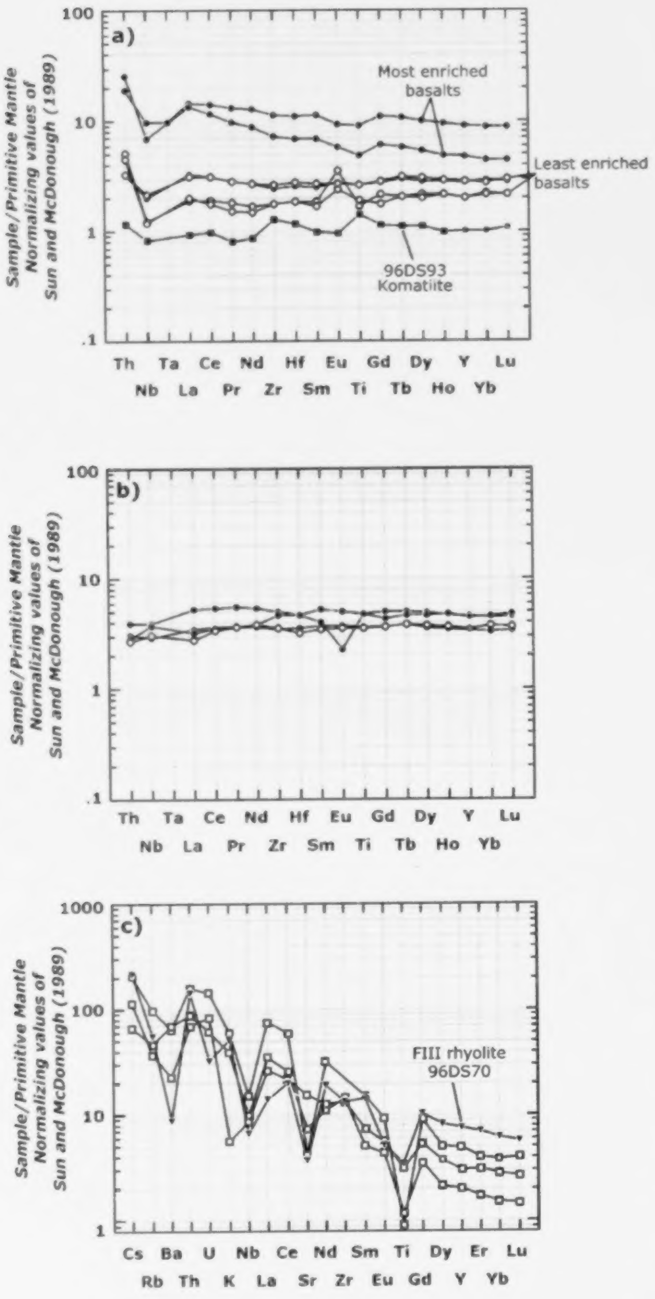


Figure 8. Multi-element plots for volcanic rocks of the Ponask Lake greenstone belt. (a) most enriched basalt (solid circles), least enriched basalt (open circle) and komatiite (squares). (b) depleted basalt of the Sachigo assemblage (solid circles) and Pierce assemblage (open circles). (c) FI rhyolite (open squares), FIII rhyolite (triangles).

Geochronology indicates closely grouped ages in the range of 2865 to 2857 Ma for biotite tonalite and supracrustal rocks of the Ponask Lake greenstone belt. The biotite tonalite at Ponask Lake has an age of 2863 Ma (No. 5 of Table 3) and detrital grains in two samples of conglomerate have uniform populations of zircon with ages of 2864 and 2865 Ma (Nos. 1 and 7 of Table 3). The ages of the tonalite and conglomerate are nearly identical within the limits of error assigned to the analyses and support the interpretation that the detritus in the conglomerate is derived from the tonalite. A quartz porphyry intrusion in volcanic rocks at Ponask Lake has an age of 2857 Ma (No. 8 of Table 3).

Sample 96DS93 of komatiite from Sachigo Lake is composed of talc+chlorite+magnesite with accessory magnetite, chromite, pyrrhotite and pentlandite. A thin section shows relict olivine or pyroxene grains surrounded by grains of magnetite. The komatiite has 26.8% MgO and high Ni (1000 ppm) and Cr (3250 ppm). The primitive-mantle-normalized Gd/Yb ratio of 1.1 and an Al_2O_3/TiO_2 ratio of 17 is representative of Al-depleted komatiite although transitional to Al-undepleted or Munro-type komatiite (Al_2O_3/TiO_2=20; Arndt 1994). Where normalized to primitive mantle (Figure 8a), the komatiite has element concentrations approximately equal to primitive mantle values, although slightly enriched in Th, Zr and Ti and possibly Hf, and is depleted in Nb. Rare earth elements (REE) show a trend of very slight depletion in light rare earth elements (LREE) except for La and Ce, which are enriched to approximately primitive mantle values.

At least two groups of basaltic komatiite and basalt are distinguished on the basis of enrichment or depletion in incompatible elements. The enriched suite is characterized by flat to slightly sloped trace element profiles enriched from 2 to 20 times primitive mantle values with pronounced enrichment of Th and depletion of Nb (*see* Figure 8a). The depleted basalts have trace element profiles that are flat to very slightly sloped to the left at 2 to 6 times primitive mantle values (*see* Figure 8b). The trace element profiles are somewhat concave up in the middle with depletion in Th, Nb and LREE and one sample is depleted in Eu.

The intermediate to felsic volcanic rocks at Ponask Lake are variably enriched in SiO_2 (64 to 78%) but lack the strong enrichment in alkalis found in the Stull Lake belt (discussed later). Where normalized to primitive mantle these calc-alkalic rocks show strongly sloped trace element profiles (Figure 8c). The multi-element profiles are enriched in Cs, Th, U and LREE with troughs for Rb, Ba, Nb, Sr and Ti.

Three samples are strongly enriched in LREE with Zr/Y ratios in the range of 10 to 20 at <20 ppm Y, characteristic of the FI-type rhyolites of Lesher et al. (1986). Sample 96DS70, representing the oval mass of intermediate to felsic fragmental rocks at western Ponask Lake, has a flatter distribution of REE with a negative Eu anomaly and a Zr/Y value of 4.3 at 35 ppm Y. This rock is characteristic of the FIIIa type rhyolite of Lesher et al. (1986) and possibly represents a high level magma chamber capable of providing heat to generate sea-floor volcanogenic massive sulphide (VMS) deposits.

The units of conglomerate, marble and komatiite are interpreted to represent platform-type sequences of Thurston and Chivers (1990) and suggest that northern and probably basal parts of Pierce and Sachigo assemblages were deposited in a continental margin environment. The occurrence of komatiites with thin sedimentary sequences overlying tonalite provides geologic evidence that the komatiite erupted through continental crust. Hence, crustal contamination may account for the enrichment in Th, La and Ce and depletion in Nb in the komatiite. Likewise, although low Nb and high Th are characteristics of basalts erupted in volcanic arcs (Wood et al. 1979), crustal material such as tonalite that has originated in continental or oceanic magmatic arcs is also enriched in Th and LREE and depleted in Nb. Hence, the enriched trace element characteristics of the enriched suite of basalts provide an indication that these magmas erupted in a continental environment where they assimilated crustal material.

The Th and LREE-depleted basalts are best explained as having originated from a mantle source transitional between primitive mantle and depleted upper mantle. In other words, the slight Th and LREE depletion of these rocks has been inherited from the source. The depleted basalts evidently erupted in an oceanic environment such as an oceanic plateau and were not contaminated by continental crust.

In general, the calc-alkalic suite shows strong depletion in Nb and Ti characteristic of volcanic sequences developed in volcanic arc environments. Somewhat variable depletion in Rb, Ba, Sr, K and Eu may be an indication of redistribution of these elements by alteration or that feldspars have been removed from the magmas by fractionation.

The Sachigo and Pierce assemblages represent a composite of continental platform sediments and mafic to ultramafic volcanic sequences that are both enriched and depleted in incompatible elements. The Sachigo assemblage also hosts strongly enriched felsic volcanic strata. Evidently these supracrustal strata developed in a variety of geologic environments including continental platform, continental margin arc and oceanic settings and have possibly been tectonically juxtaposed across unrecognized faults.

STULL–SWAN LAKES GREENSTONE BELT

The Stull–Swan lakes greenstone belt is part of a complexly curved and variably bifurcated and anastomosed chain of greenstone belts. The chain of greenstone belts extends from Oxford Lake through Knee Lake and Gods Lake, Manitoba to Stull Lake, Ontario and hence, east-southeast through Swan Lake to Big Trout Lake over a distance of 400 km. The greenstone belt underlies approximately 7% of the northern Superior area and attains a maximum width of 12 km at Stull Lake. The belt is composed of several undated basaltic assemblages and Neoarchean, calc-alkalic to alkalic volcanic and clastic sedimentary assemblages.

The Stull–Swan lakes greenstone belt at Stull Lake can be subdivided into 3 structural panels each on the order of 2 to 8 km thick. These include bounding panels of mafic volcanic rocks and an internal panel of interleaved felsic volcanic and clastic sedimentary rocks.

Edmund Lake Assemblage

The Edmund Lake assemblage occurs at the south side of the Stull–Wunnummin fault. It extends from Edmund Lake, Manitoba to south of Little Stull Lake and tapers through the northern part of Stull Lake. The Edmund Lake assemblage is represented by pillowed to massive, aphyric basalt flows and associated gabbro dikes and sills. Although undated, these basalt sequences are intruded by 2734 Ma tonalite at Margaret Lake, Manitoba (No.18 of Table 1).

Corkery and Heaman (1998) noted that basalts of the Edmund Lake assemblage are tholeiitic and have MORB (mid ocean ridge basalt)-like trace element characteristics. Samples have REEs enriched to 10x chondritic values and can be slightly LREE depleted. One sample of the Edmund Lake assemblage from northern Stull Lake is high-iron tholeiite (Figure 6b) and is classified as belonging to the low-K series (Figure 7b). Trace elements in this sample are enriched from 3 to 5 times primitive mantle values and are slightly depleted in Th and Nb with Th<Nb<La (00DS199 of Figure 9b).

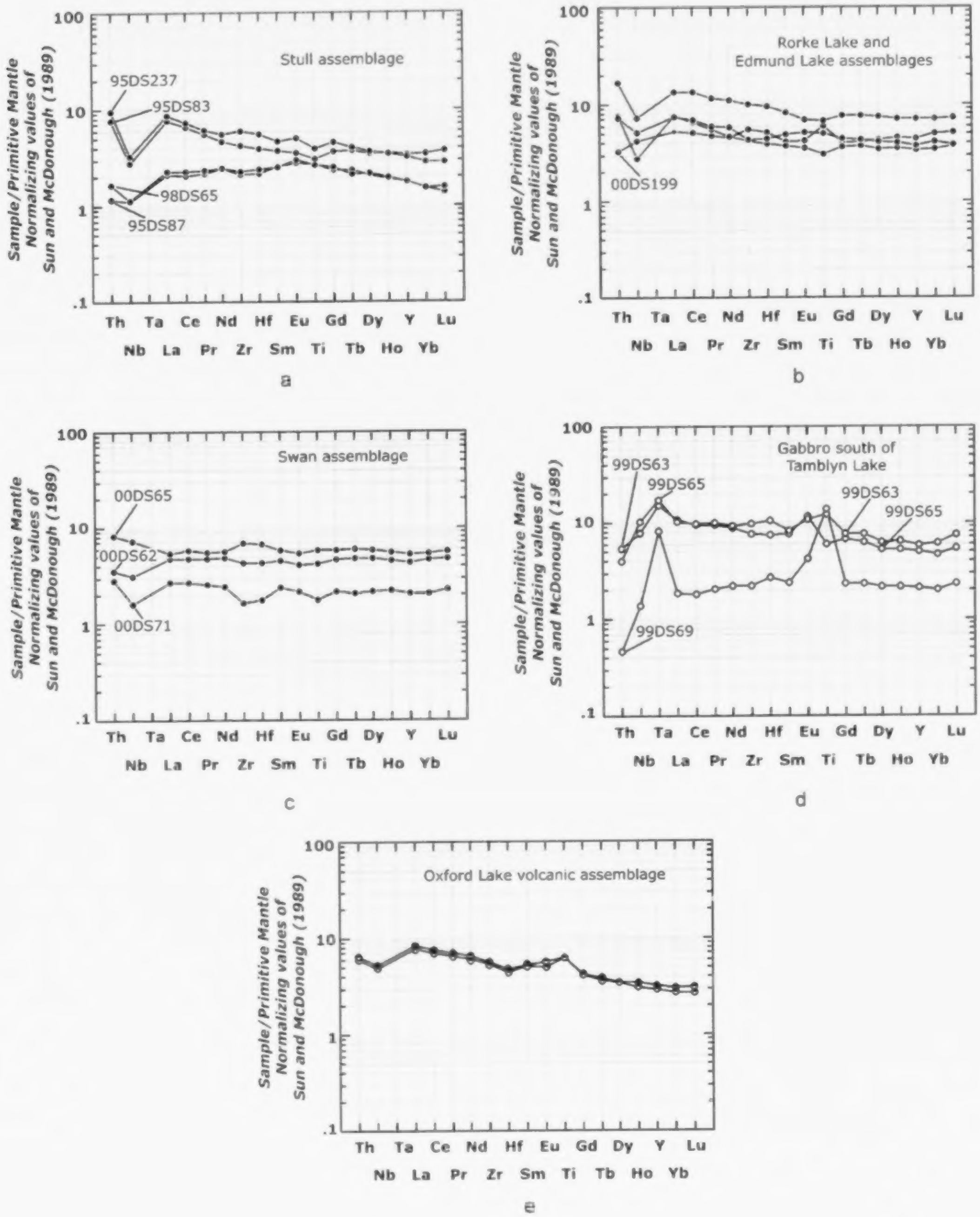


Figure 9. Multi-element plots for mafic volcanic rocks of the Stull–Swan lakes greenstone belt.

Rorke Lake Assemblage

The Rorke Lake assemblage occurs mainly on the north side of the Stull–Wunnummin fault in the area of Kistigan Lake and Rorke Lake (see map in back pocket). The Rorke Lake assemblage extends eastward in the Ellard Lake greenstone belt and southeasterly along the north side of the Stull–Swan lakes greenstone belt at Stull Lake. A thin unit representative of the Rorke Lake assemblage occurs south of the Edmund Lake assemblage at north central Stull Lake.

Variably aphyric to hornblende- and plagioclase-phyric basalt flows with minor associated volcaniclastic and sedimentary rocks comprise the Rorke Lake assemblage. The undated basalt sequences are intruded by 2715 Ma hornblende tonalite at Kistigan Lake (No. 20 of Table 3). Corkery and Heaman (1998) noted that the basalts of the Rorke Lake assemblage are enriched in incompatible trace elements and have Th>La>Nb on primitive-mantle-normalized trace element plots. Three samples from the present area are mainly high-iron tholeiite (Figure 6b) and belong to the low-K series (Figure 7b). Primitive-mantle-normalized trace element profiles for the three samples of the Rorke Lake assemblage (excluding sample 00DS199, which belongs to the Edmund Lake assemblage) are sloped from left to right and generally have Th>La>Nb (Figure 9b).

Stull Assemblage

The Stull assemblage extends along the south side of the Stull–Swan lakes greenstone belt from Stull Lake to the area of Blackbear Lake. Eastern parts of this assemblage are highly sheared by the Stull–Wunnummin fault. The Stull assemblage attains a thickness of 3 km at Stull Lake and is composed of north-younging pillowed mafic lavas with associated chert and fine-grained sedimentary units.

Although contacts are not well exposed, the Stull assemblage appears to be intruded by 2848 Ma tonalite and tonalite gneiss at southern Stull Lake (No. 3 of Table 3). Mafic volcanic rocks of the Stull assemblage are mainly high-magnesium tholeiite and calc-alkalic basalt (Figure 6b) belonging to the low-K series (Figure 7b). Samples of the Stull assemblage fall into two groups on the basis of primitive-mantle-normalized trace element characteristics. The profiles of samples 95DS237 and 95DS83 are sloped from left to right with Th>La>Nb whereas the profiles of samples 98DS65 and 95DS87 are concave up in the middle with Th≈Nb<La (Figure 9a).

Swan Assemblage

The Swan assemblage comprises several narrow and highly metamorphosed greenstone slivers in the eastern Stull–Swan lakes greenstone belt (see map in back pocket). The undated greenstone slivers are made up of amphibole gneisses that probably originated as mafic lava flows and have subsequently become complexly interleaved with other supracrustal and plutonic rocks. Three samples of the Swan assemblage are basaltic komatiite and calc-alkalic basalt (Figure 6b) and belong to the low-K series (Figure 7b). Trace element profiles are fairly flat at 2 to 8 times primitive mantle values (Figure 9c). Sample 00DS62 is slightly depleted in Th whereas 00DS65 and 00DS71 are slightly enriched in Th compared to other elements.

Tamblyn Lake Gabbro

A crescentic body of strongly foliated coarse-grained hornblende gabbro occurs south of Tamblyn Lake in the Stull–Swan lakes greenstone belt. The Tamblyn Lake gabbro ranges compositionally from high-iron tholeiite to tholeiitic rhyolite (Figure 6b) with the rhyolitic compositions reflecting local plagioclase-rich phases within the gabbro body. The gabbro has low SiO_2 and K_2O (Figure 7b).

Trace element profiles for the Tamblyn Lake gabbro are overall flat and strongly depleted in Th and Nb relative to La (Figure 9d) with evidence of fractionation. With increasing SiO_2 and Al_2O_3, trace element abundances increase from 2 to 10 times primitive mantle and shift from slight LREE depletion to slight LREE enrichment. All samples have Lu>Yb, which is an indication that olivine has fractionated from the magma. Ta and Ti are strongly enriched in the SiO_2-poor sample (99DS69) and weakly enriched or depleted in the SiO_2-rich samples (99DS63 and 99DS65). These characteristics suggest that a Ti-oxide or titanite in which Ta and Ti are strongly partitioned (Rollinson 1996) have fractionated from magma and accumulated mainly in the SiO_2-poor phases of the gabbro. Enrichment in Eu in all samples indicates that the gabbro can be a cumulate of plagioclase crystals removed from other parts of the magma chamber. Strong depletion in Th is interpreted to reflect the source composition of the magma and to indicate that the Tamblyn Lake gabbro evolved, possibly in an oceanic environment where it did not interact with older crust.

Pillowed Mafic Flow at Central Stull Lake

Remarkably well-preserved pillowed mafic volcanic rocks are exposed on an island in central Stull Lake. The pillowed volcanic rocks comprise a thin and possibly fault-bounded unit that is traced on airborne geophysical magnetometer maps (E. Downie, Geologist, Wolfden Resources, personal communication, 2000) though central Stull Lake (*see* Stone, Hallé and Pufahl 2001b).

Two samples of the fresh pillowed volcanic rocks are high-iron tholeiite (Figure 6b) of the low-K series (Figure 7b). The pillowed mafic flows are LREE-enriched with La>Th>Nb and are also enriched in Ti and somewhat depleted in HREE relative to other trace elements (Figure 9e). Enrichment in Ti and slight depletion in HREE are consistent with the magma having separated from a residuum that contained a small amount of garnet, with the implication that melt extraction occurred at depths greater than that from which other basalt sequences were derived. On the discrimination diagram of Pearce and Cann (1973), samples of the pillowed mafic flow at central Stull Lake plot in the field of within-plate basalts, distinct from all other basalts that scatter between fields of volcanic arcs and ocean ridges. The magma erupted in the pillowed flow at central Stull Lake has possibly assimilated crustal material, giving rise to enrichment in Th and LREE. This flow is likely much younger than other mafic sequences at Stull Lake and is interpreted to belong to the Oxford Lake volcanic assemblage.

Summary of Mafic Assemblages in the Stull–Swan Lakes Greenstone Belt

Excluding the fresh, pillowed volcanic unit at central Stull Lake, geochemistry reveals at least two distinct groups of basalt in the Stull–Swan lakes greenstone belt. These are depleted basalts representative of the Edmund Lake assemblage and part of the Stull assemblage and enriched basalts within the Rorke Lake and part of the Stull Lake assemblage. The Tamblyn Lake gabbro and the Swan assemblage appear to be chemically similar to the depleted basalts, although minor Th-enrichment occurs in rocks of the Swan assemblage.

The depleted and enriched mafic lavas in the north panel at Stull Lake (excluding the pillowed mafic flow at central Stull Lake) are chemically comparable to basalts at Little Stull Lake and Edmund Lake, Manitoba (Corkery and Heaman 1998; Corkery and Skulski 1998). These basalt sequences are correlated with the 2.83 Ga Hayes River assemblage. The depleted basalts are interpreted to represent magmas derived from a mantle source that was depleted in Th, Nb and LREE and erupted in an oceanic environment - possibly an oceanic plateau based on limited La/Yb vs. Th/Ta data (*see* Condie 1997). The enriched basalts have characteristics of lavas erupted in volcanic arcs (Th and LREE enrichment and depletion in Nb; Pearce 1996) or else mafic lavas that have been contaminated with crustal material. ε_{ND} values are somewhat lower for enriched basalts than for depleted basalts and were interpreted to indicate that the enriched basalts could have erupted in a continental margin arc where they assimilated crustal material (Skulski et al. 2000). In contrast, the Nd-isotope systematics suggest a juvenile oceanic source for the depleted basalts and imply at least two chemically distinct basalt sequences within the Stull–Swan lakes greenstone belt.

The Stull assemblage is chemically similar to basalt sequences in the Sachigo assemblage of the Ponask Lake greenstone belt. The Stull assemblage is bounded on the north by a major fault and on the south by 2.85 Ga tonalitic rocks (Nos. 3, 4 of Table 3), which are similar in age to the Sachigo assemblage. On the basis of this geochemical, structural and geochronological evidence, the Stull assemblage is interpreted to be part of the Island Lake terrane. The Stull–Wunnummin fault marks the north boundary of the Island Lake terrane.

Oxford Lake Volcanic Assemblage

Intermediate to felsic volcanic rocks occur within three elongate units in the central Stull–Swan lakes greenstone belt at Stull Lake. The upper or northernmost unit can be traced east from Twin Lakes for 40 km to Stull Lake and is composed of coarse, white to tan-brown volcanic breccias, bedded tuffs and flows interbedded with pale green mafic flows. On an island in Stull Lake, a zone of carbonate+quartz alteration possibly representing a metamorphosed iron formation unit is overlain southward by a fresh, pillowed mafic flow (described above) and felsic volcanic breccia. Eastward, these volcanic rocks extend as a tapered unit into the heterogeneous Gilleran Lake intrusion (*see* map in back pocket). The Gilleran Lake intrusion varies compositionally from gabbro and diorite at the margins to tonalite and granodiorite at the core. The volcanic rocks at Stull Lake show mineral assemblages characteristic of greenschist-facies metamorphism. For example, the pillowed mafic flow within the sequence has the mineral assemblage chlorite+albite+epidote+calcite+magnetite.

The middle unit of intermediate to felsic volcanic rocks extends east-southeast through central Stull Lake locally attaining a width of 1 km. It is separated from the northernmost unit of volcanic breccias by pebbly sandstone and is unconformably overlain by conglomerate to the south. The unit is composed of volcanic tuffs and breccias in the basal northern section giving way southward to distinct grey lava flows and breccias containing feldspar and amphibole phenocrysts. Locally, the flows show ribbon lava textures suggestive of subaerial eruption.

The lower (southernmost) unit of intermediate to felsic volcanic rocks at Stull Lake is strongly deformed by splays of the Stull–Wunnummin fault and interleaved with sedimentary rocks on the east shore of Stull Lake. The lower unit is composed of tuffs, flows and breccias and appears to lack the porphyritic rocks that are characteristic of the central unit.

Intermediate to felsic volcanic rocks occur at several localities within narrow greenstone slivers in the area of Swan Lake. These units are typically composed of tuffs and breccias with minor components of siliceous rhyolite flows and quartz+feldspar porphyry at Swan Lake. Intermediate to felsic volcanic rocks in the eastern Stull–Swan lakes greenstone belt are strongly foliated and appear to be tectonically mixed with mafic volcanic and sedimentary rocks.

Corkery and Skulski (1998) described sequences of intermediate to felsic tuffs and breccias and associated epiclastic sedimentary rocks at Little Stull Lake and Rorke Lake, Manitoba (*see* map in back pocket), which they included with the Oxford Lake volcanic subgroup or Oxford Lake volcanic assemblage. Broadly, the intermediate to felsic volcanic rocks of the Oxford Lake volcanic assemblage are calc-alkalic grading to alkalic compositions associated with feldspar- and hornblende-phyric lavas. The available geochronology supports the interpretation that the intermediate to felsic volcanic sequences of the Stull–Swan lakes greenstone belt are part of the Oxford Lake volcanic assemblage. Analyses 10, 12, 13, 14, 19 and 26 of Table 3 are derived from samples of intermediate to felsic volcanic rocks in various parts of the Stull–Swan lakes greenstone belt and show ages in the range of 2726 to 2717 Ma. These ages are distinctly younger than those of the Sachigo and Hayes River assemblages and are consistent with the ages of Oxford Lake volcanic rocks in other parts of the northern Superior Province (compare the above data with Table 1).

Volcanic rocks of the Oxford Lake volcanic assemblage in the Stull–Swan lakes greenstone belt that were mapped in the field as intermediate to felsic in composition are shown on the basis of major element geochemistry to range from basalt and andesite through dacite and rhyolite (Figures 6c, d). Most rocks of basalt to andesite composition are derived from the upper unit. One sample from the upper unit at Stull Lake is a high-magnesium tholeiitic basalt and a pillowed flow at central Stull Lake (the geochemistry of this flow is discussed above with mafic volcanic sequences) is a high-iron tholeiitic basalt. Other rocks of basalt to andesite composition in the upper unit at Stull Lake (shown by solid dots in Figure 6c) are pale green to grey massive flows or breccias and none exhibit pillows.

Volcanic rocks from all units of the Oxford Lake volcanic assemblage are distinguished by high alkalis (typically 1 to 5 weight percent K_2O) and scatter through the medium- to high-K fields on the K_2O-SiO_2 diagram (Figures 7c, d). Samples showing feldspar and/or amphibole phenocrysts tend to have the highest concentration of alkali oxides. These and samples of the Gilleran Lake intrusion are classified as basaltic trachyandesite, trachyandesite and trachyte using the classification of Le Maitre (1989; Figure B13).

For purposes of geochemical description, volcanic rocks of the Oxford Lake volcanic assemblage are subdivided geographically according to the upper, middle and lower units at Stull Lake, the Swan Lake area and the Gilleran Lake intrusion. Samples from the northern unit at Stull Lake are further subdivided on the basis of major element geochemistry into groups ranging compositionally from basalt to andesite and from dacite to rhyolite.

The primitive-mantle-normalized multi-element profiles of all samples of the Oxford Lake volcanic assemblage are steeply sloped to the right due to strong enrichment in the incompatible elements (Figures 10a-f). Element enrichment varies significantly from sample to sample with the result that trace element profiles are shifted upward by varying distances and plot within a broad band. Deep troughs for Nb and Ti are common to all profiles and provide an indication that the volcanic rocks of the Oxford Lake volcanic assemblage developed in magmatic arcs. The concentrations of large-ion lithophile elements (LILE) (Cs, Rb, Ba, Sr) as well as Th and U scatter somewhat, probably indicating that these elements are mobile and have been affected by alteration. In spite of alteration, Th>La>Nb in most samples, which is consistent with magmas generated in volcanic arcs as well as magmas that have assimilated crustal material. Zr tends to be depleted in some samples including those that are strongly enriched in other trace elements. This is interpreted to result from fractionation of zircon or hornblende in the chemically evolved samples.

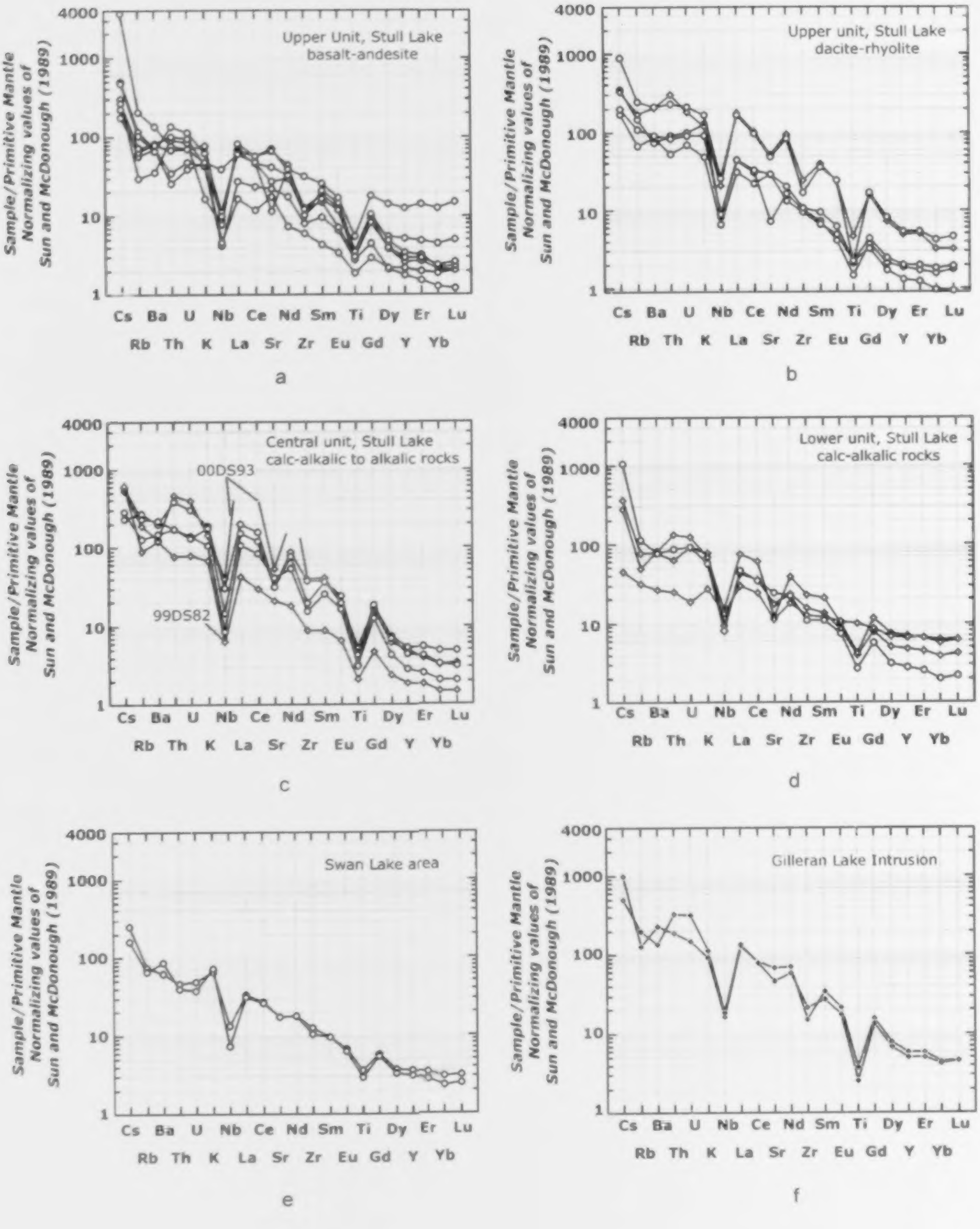


Figure 10. Multi-element plots for intermediate to felsic volcanic rocks of the Stull–Swan lakes greenstone belt.

Minor chemical variations occur between the various subdivisions of the Oxford Lake Group. For example, in the northern unit at Stull Lake, rocks of dacite to rhyolite composition tend to be slightly enriched in LILE and LREE compared to basalt and andesite. The central unit at Stull Lake shows evidence of evolution in the composition of erupted lavas (*see* Figure 10c). Sample 99DS82 of calc-alkalic breccia from near the northern base of the unit is the least enriched in trace elements, whereas sample 00DS93 of an alkalic tuff from the top of the unit is highly enriched to the extent that LREE and Nd could not be measured by the ICP-MS. On the basis of chemical similarity with erupted lavas, the Gilleran Lake intrusion is interpreted to be a subvolcanic intrusion of the Oxford Lake volcanic assemblage.

In comparison with older mafic volcanic sequences, the Oxford Lake volcanic assemblage is chemically more variable. The Oxford Lake volcanic assemblage has variably higher concentrations of alkali elements including K_2O (*see* Figure 7), LILE and LREE and a greater range of La/Yb ratios. Rocks of basalt to andesite composition in the upper unit of the Oxford Lake volcanic assemblage at Stull Lake are Zr-enriched and plot within the field of volcanic arc basalts on the discrimination diagram of Pierce and Cann (1973). In contrast, older basalts of the various mafic assemblages are Zr-poor and plot in fields transitional from volcanic arc to oceanic basalts.

Clastic Sedimentary Sequences (Oxford Lake Sedimentary Assemblage and Cross Lake Assemblage)

Coarse clastic sedimentary rocks extend 30 km between Twin Lakes and northern Stull Lake and are the predominant rock type between the upper and central units of the Oxford Lake volcanic assemblage at Stull Lake. Coarse clastic sedimentary rocks unconformably overly the central unit of the Oxford Lake volcanic assemblage and grade southward to thin-bedded wacke sequences at southern Stull Lake. Thin sequences (typically up to 2 km wide) of fine clastic sedimentary rocks occur at Rapson Bay of Stull Lake and in the areas of Meston, Tamblyn, Swan and Blackbear lakes in the eastern parts of the greenstone belt (*see* map in back pocket).

Coarse clastic sedimentary rocks at Stull Lake include conglomerate, pebbly sandstone and sandstone. Clasts within the conglomerate include hornblende tonalite (the hornblende is altered to actinolite) and fragments representative of the Oxford Lake volcanic assemblage and various mafic assemblages. The conglomerate grades southward at Stull Lake through pebbly sandstone to arkosic sandstone. The sandstone is tan-brown and feldspar-rich with less than 10 percent sericite and chlorite and often has carbonate developed within the matrix of feldspar and quartz grains. The sedimentary sequences are massive to thick bedded and locally cross-bedded and graded. The tops of graded beds are to the south at most localities.

Well-bedded wacke-siltstone sequences occur south of the conglomerate and sandstone at southern Stull Lake. The fine-grained sedimentary rocks young southward and appear to be in fault contact with the Stull assemblage. Wacke-siltstone sequences and units of the Oxford Lake volcanic assemblage also extend southeasterly through Rapson Bay of Stull Lake adjacent to the northern panel of the Rorke Lake assemblage and occur at scattered localities in the eastern Stull–Swan lakes greenstone belt. The sedimentary rocks within these units lack recognizable coarse clastic detritus and are composed of alternating thin beds of wacke and siltstone. The wacke contains more quartz and mafic minerals (typically 30% biotite) than sandstone of similar grain size at Stull Lake. Typical of metamorphosed supracrustal sequences in thin greenstone slivers, the sedimentary rocks in the eastern Stull–Swan lakes greenstone belt are foliated, folded and locally migmatized and are cut by felsic dikes.

Corkery, Skulski and Whalen (1997) described two sequences of sedimentary rocks at Little Stull Lake. These include 1) wacke turbidites and argillite interbedded with felsic tuffs and volcanic rocks of the volcanic subgroup of the Oxford Lake Group and 2) coarse sandstone and conglomerate. On the basis of a reinterpretation of sedimentary sequences in northern Manitoba (T. Corkery, Geologist, Manitoba Energy and Mines, personal communication, 2002; *see* "Subdivisions of Supracrustal and Plutonic Rocks: Manitoba"), the fine-grained wacke sequences and coarse clastic sequences are assigned to the Oxford Lake sedimentary assemblage and Cross Lake assemblage, respectively. The two-fold subdivision of sedimentary sequences at Little Stull Lake can be extended broadly through the present area and implies two depositional environments. The wacke-siltstone units are characteristic of sedimentary sequences deposited by turbidity currents in a deep subaqueous and probably marine environment. In contrast, the coarse clastic sedimentary rocks show evidence of subaerial deposition within fluvial to alluvial fans on the flanks of pull-apart basins along the Stull–Wunnummin fault.

The relative ages and stratigraphic relations between the subaqueous and subaerial sedimentary sequences are unclear and possibly variable. On the one hand, south-younging strata at Stull Lake imply that the deep subaqueous sediments (wacke-siltstone sequences at southern Stull Lake) overlie the subaerial sediments (conglomerate at central Stull Lake). This is consistent with a fining-upward sequence developed in a deepening fault-bounded basin. On the other hand, several lines of evidence suggest that the wacke-siltstone sequences developed in depositional environments distinct from that of the subaerial sediments and at an earlier time. For example, the wacke-siltstone sequences at southern Stull Lake are cut by splays of the Stull–Wunnummin fault and have possibly been tectonically juxtaposed with coarse clastic rocks to the north. Further, the youngest detrital zircon grains from samples associated with wacke-siltstone sequences are 2723 and 2719 Ma, whereas those from sandstone sequences are 2713 and 2709 Ma (*see* analyses 15 to 18 of Table 3). Although inconclusive, the geochronology provides evidence that the wacke-siltstone sequences can be older than the conglomerate-sandstone sequences. Skulski et al. (2000), citing isotopic evidence, suggested that volcanic rocks of the Oxford Lake volcanic assemblage evolved from early (>2.72 Ga) juvenile calc-alkalic phases to late (2.71 Ga) crustally contaminated alkalic phases. This evolution occurred as volcanism changed from an oceanic to continental environment and the late alkalic magmas passed through and assimilated crustal material. Accordingly, marine sedimentary sequences such as occur at Rapson Bay and Swan Lake in association with calc-alkalic phases of the Oxford Lake volcanic assemblage, can be older than coarse fluvial-alluvial sequences overlying alkalic phases of the Oxford Lake volcanic assemblage at Stull Lake. Southern Stull Lake may represent a locality where a large basin developed within which coarse fluvial-alluvial sequences are overlain by fine wacke-siltstone sequences of the Cross Lake assemblage or else where coarse sequences have been tectonically interleaved with the fine sediments of the Oxford Lake sedimentary assemblage.

ELLARD LAKE GREENSTONE BELT

The Ellard Lake greenstone belt extends easterly from Little Stull Lake through Ellard Lake and Foster Lake and hence, east-southeasterly over a total length of 130 km (*see* map in back pocket). The Ellard Lake greenstone belt merges with the Stull–Swan lakes greenstone belt at its western end and appears to taper within felsic plutonic rocks at its eastern terminus east of the present area. The belt underlies approximately 3% of the area and attains a width of 5 km, locally bulging around intrusive oval plutons of the sanukitoid suite such as at Ellard Lake. The Ellard Lake greenstone belt is composed of two mafic assemblages (Rorke Lake and Ellard), the intermediate to felsic Oxford Lake volcanic assemblage and sedimentary sequences representative of the Cross Lake and Oxford Lake sedimentary assemblages.

Rorke Lake Assemblage

The Rorke Lake assemblage comprises the western part of the Ellard Lake greenstone belt and extends west from Ellard Lake to Little Stull Lake attaining a width of 3 km (*see* map in back pocket). The Rorke Lake assemblage is poorly exposed and a few outcrops east of Echoing Lake show pillowed to massive mafic volcanic flows. No samples were taken for chemical analysis.

Ellard Assemblage

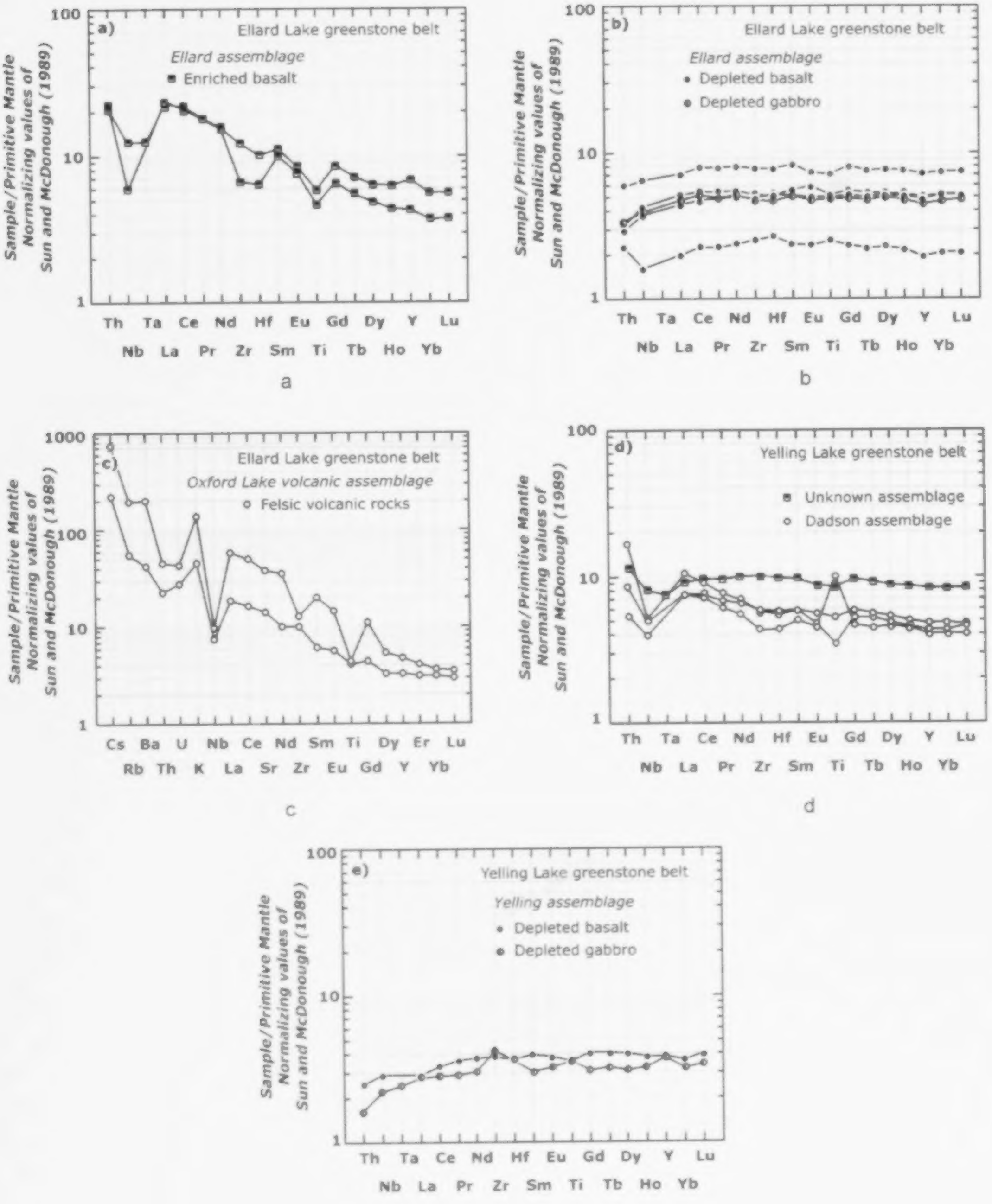
Between Ellard Lake and Foster Lake, the Ellard Lake greenstone belt is broadly synclinal with inward-younging mafic volcanic rocks at the base of the supracrustal section. The mafic sequences are represented by approximately equal proportions of massive and pillowed lavas with a subordinate component of medium- to coarse-grained gabbro sills. Feldspar-phyric mafic flows occur locally. Thin units (up to 10 m thick) of wacke and chert occur between lava flows at scattered localities.

Mafic sequences of the Ellard assemblage are metamorphosed to amphibolite facies. For example, a pillowed mafic flow at Foster Lake has a mineral assemblage of hornblende+biotite+plagioclase+ zoisite+titanite. Within thin greenstone slivers east of Foster Lake, deformation by the South Kenyon fault has obliterated primary features and the mafic sequences are transformed to amphibole gneisses.

Mafic sequences of the Ellard assemblage young north from an unexposed contact with 2734 Ma biotite tonalite (analysis 24 of Table 3) 15 km southeast of Ellard Lake. Although the contact can represent an unconformity (in which case the mafic sequences of the Ellard Lake greenstone belt are younger than 2734 Ma) other evidence suggests that the contact is faulted. The results of regional Al-in-hornblende barometry (discussed later) show that crystallization pressures are higher for plutonic rocks south of the Ellard Lake greenstone belt than in plutonic rocks of other areas, particularly north of the belt. The results are interpreted to indicate that the Ellard Lake greenstone belt has been thrust underneath the plutonic rocks south of the belt and has caused uplift of the plutonic complex south of the greenstone belt (*see* Figure 4). In this case, the southern contact of the Ellard Lake greenstone belt is marked by an unexposed fault and the structural facing relationships of the mafic sequences to the biotite tonalite provide no insight on the age of the mafic sequences.

Mafic sequences of the Ellard assemblage range compositionally from basaltic komatiite to high-iron tholeiite (Figure 6e). The majority of samples plot within the low-K field in the K_2O-SiO_2 diagram of Le Maitre (1989) although basalts enriched in Th and LREE have somewhat elevated K_2O and are classified as medium-K (Figure 7e).

Mafic rocks of the Ellard assemblage are subdivided into enriched and depleted varieties on the basis of multi-element characteristics. The enriched variety (Figure 11a) is strongly enriched in Th and LREE with depletion in Nb, Ta and Ti and slight depletion in Zr and Hf. The depleted variety, including a gabbro sill at Foster Lake, has an overall flat multi-element profile with depletion in Th, Nb and LREE and Th<Nb<La, although one sample lacks the Th-depletion (Figure 11b). On the Th-Hf-Nb discrimination diagram of Wood et al. (1979; not shown) the enriched and depleted samples plot within the fields of volcanic arc basalts and mid ocean ridge basalts, respectively. As discussed previously (see "Summary of Mafic Assemblages in the Stull–Swan Lakes Greenstone Belt") the enriched basalts could have alternatively developed through crustal contamination.



a

b

c

d

e

Figure 11. Multi-element plots for volcanic rocks of the Ellard Lake and Yelling Lake greenstone belts.

Oxford Lake Volcanic Assemblage

Intermediate to felsic volcanic sequences attain a thickness of up to 1 km within a crescentic unit at Ellard Lake and a bifurcated unit at the central axis of the Ellard Lake greenstone belt at Sherman Lake (*see* map in back pocket). Volcanic breccias and tuffs with a lesser component of flows are recognized in broad parts of the felsic units whereas narrow extensions of the felsic units are gneissic and probably of tuffaceous and sedimentary origin. A quartz+feldspar porphyry sill is traced three kilometres west of Sherman Lake. Lenses of wacke occur at scattered localities and iron formation is noted at one outcrop within the sequence at Sherman Lake.

A fragmental felsic volcanic rock 1 km south of Ellard Lake and a quartz+feldspar porphyry sill in mafic rocks at Foster Lake have ages of 2732 and 2730 Ma, respectively (Nos. 11 and 25 of Table 3). The sill has an older inherited zircon fraction with an age of 2736 Ma. These ages are a few Ma older than the dates for volcanic rocks of the Oxford Lake volcanic assemblage in other northwestern greenstone belts but are somewhat younger than a 2734 Ma tonalite stock at Margaret Lake (No. 18 of Table 1). Corkery and Heaman (1998) concluded on the basis of geochemistry that the Margaret Lake stock could represent a subvolcanic intrusion associated with the Oxford Lake volcanic assemblage. Likewise, the 2730 to 2732 Ma intrusive and extrusive rocks of the Ellard Lake greenstone belt are interpreted to represent the early stages of volcanism associated with the Oxford Lake volcanic assemblage.

Samples of a massive, brown-weathered rock possibly representing a volcanic flow and a grey volcanic breccia were sampled for geochemistry. These samples are calc-alkalic basalt and dacite, respectively (open circles of Figure 6e). The basalt has high-K whereas the dacite plots within the medium-K field on the K_2O-SiO_2 diagram (Figure 7e). Multi-element profiles are strongly sloped to the right (Figure 11c) due to enrichment in LILE and LREE. Like volcanic rocks of the Oxford Lake volcanic assemblage from other greenstone belts, the multi-element profiles have troughs for Nb and Ti. The Ellard Lake samples have somewhat lower Th and U than volcanic rocks of the Oxford Lake volcanic assemblage in other belts.

Clastic Sedimentary Sequences (Cross Lake Assemblage and Oxford Lake Sedimentary Assemblage)

Coarse polymictic conglomerate is exposed in rare outcrops between Ellard Lake and the Echoing River. The logs of exploration diamond-drill holes in this area (Assessment Files, Resident Geologist's Office, Red Lake) also describe intersections of sedimentary rocks including conglomerate. The conglomerate contains a mixture of volcanic and plutonic clasts representative of the Cross Lake assemblage and extends southeasterly giving way to thin-bedded chlorite+biotite schists representative of wacke-siltstone sequences south of Ellard Lake (*see* map in back pocket). The fine-grained sedimentary rocks (Oxford Lake sedimentary assemblage) appear to abut against 2734 Ma biotite tonalite to the south and volcanic sequences of the Ellard Lake and Oxford Lake volcanic assemblages to the north, however the contacts are not exposed and possibly faulted at this locality.

Strongly foliated conglomerate occurs in the riverbed east of the outlet of Sherman Lake. The conglomerate contains stretched felsic pebbles probably representative of mixed intrusive and volcanic protoliths occurring in a dark chloritic matrix. The conglomerate unit at this locality is less than 100 m wide and occurs between units of mafic and felsic volcanic rocks possibly at the synclinal axis of the belt. Sedimentary units of the Ellard Lake greenstone belt have not been dated, however some units are characterized by a low grade of metamorphism and a high stratigraphic position relative to volcanic rocks. These metamorphic and stratigraphic conditions are consistent with what is found regionally in rocks of the Cross Lake and Oxford Lake sedimentary assemblages.

YELLING LAKE GREENSTONE BELT

The Yelling Lake greenstone belt extends discontinuously east-southeasterly for 200 km adjacent to the North Kenyon fault, over which distance it rarely attains a width of more than 2 km. It is the smallest greenstone belt, underlying approximately 1.3% of the northern Superior area. To the west, the Yelling Lake greenstone belt merges with other greenstone sequences at Red Cross Lake, Manitoba and to the east it bifurcates into greenstone slivers that dissipate among felsic plutonic rocks 20 km east of the northern Superior area (Bennett and Riley 1969). Although poorly exposed, narrow and deformed, the Yelling Lake greenstone belt contains diverse supracrustal sequences. Three mafic volcanic assemblages as well as rocks representative of the Oxford Lake volcanic and sedimentary assemblages are present.

Dadson Assemblage

The Dadson assemblage represents a series of narrow, east-trending greenstone slivers in the Dadson Lake area (see map in back pocket). The greenstone slivers are composed primarily of black amphibole gneisses although pillowed volcanic flows and coarse-grained gabbro bodies are observed locally. Highly deformed felsic tuffs and intermediate volcanic rocks occur at Dadson Lake. The felsic tuff within mafic sequences at Dadson Lake is 2838 Ma old and has an inherited component of 2847 Ma zircon (analysis 30 of Table 3). The Dadson assemblage has evidently erupted at the same time as the Hayes River assemblage was deposited in the Knee Lake area (Nos. 5 to 7 of Table 1).

Geochemically, mafic sequences of the Dadson assemblage are high-magnesium to high-iron basalts (Figure 6f), and of the low-K series (Figure 7f). Primitive-mantle-normalized trace element profiles are sloped from left to right due to enrichment in incompatible elements (Figure 11d). The basalts of the Dadson assemblage are enriched in Th and LREE and are depleted in Nb and either depleted or enriched in Ti relative to other elements. These geochemical characteristics of the Dadson assemblage can be interpreted to indicate that eruption took place in a volcanic arc, or else the magmas assimilated older crustal material. Although the basalts plot marginally within the field of volcanic arc basalts on the Th-Hf-Nb diagram of Wood et al. (1979), assimilation of older crustal material is favoured by the structural position of the Dadson assemblage at the south margin of the North Caribou terrane.

Unknown Assemblage

A poorly exposed and highly deformed mafic volcanic unit extends southeasterly through Yelling Lake over a distance of 40 km, attaining a width of up to 2 km. To the northwest, this unit possibly joins with 2.83 Ga mafic sequences at Red Cross Lake, Manitoba (Corkery and Stone, in press), although the intervening area is not mapped. The unknown assemblage can also comprise part of the Dadson assemblage to the southeast although geochemically, the unknown assemblage is distinct from the Dadson assemblage.

A sample of mafic volcanic rock from the unknown assemblage is classified as high-iron tholeiite (Figure 6f) and of the medium-K series (Figure 7f). Trace elements are fairly uniformly enriched to about 10 times primitive mantle values (Figure 11d) although Th is somewhat more highly enriched and Nb, Ta and Ti are slightly depleted relative to other elements. Although these characteristics are broadly similar to rocks of the Dadson assemblage, further geochemical analyses are required to characterize the unknown assemblage and to determine if it can be correlated with other volcanic sequences in the Yelling Lake greenstone belt.

Yelling Assemblage

The Yelling assemblage is represented by a narrow, northerly trending arm of mafic volcanic rocks and an oval gabbro body, 20 km east of Yelling Lake. A sample of the coarse-grained gabbro has a U-Pb age of 2716 Ma (No. 32 of Table 3).

Samples of the mafic volcanic rock and the gabbro are high-magnesium tholeiite (Figure 6f) belonging to the medium-K series (Figure 7f). Trace element characteristics of the mafic volcanic sample are similar to those of the gabbro sample. Both have trace elements enriched 3 to 4 times primitive mantle values with depletion in Th, Nb, Ta and LREE and Th<Nb<La (Figure 11e). These plot within the MORB field on the Th-Hf-Nb diagram of Wood et al. (1979). On the basis of their depleted geochemical signature, the gabbro and basalt are interpreted to have crystallized in an oceanic environment at 2716 Ma. The depleted gabbro and basalt indicate that an ocean basin could have existed at the south margin of the Northern Superior Superterrane at 2716 Ma. Mafic magmatism evidently took place within the oceanic environment without interacting crustal material. The mafic rocks of the Yelling assemblage must have been obducted onto the craton margin at some time after 2716 Ma.

Oxford Lake Volcanic Assemblage

Intermediate to felsic volcanic rocks of the Oxford Lake volcanic assemblage are recognized at two principal localities, which are 20 km east of Yelling Lake and in the area of Dadson Lake. At the locality 20 km east of Yelling Lake, gneissic rocks of probable tuffaceous origin appear to be interbedded with fine clastic sedimentary rocks and extend northerly adjacent to the Yelling assemblage. Several types of stretched intermediate to felsic volcanic clasts are exposed within a layered sequence at one outcrop east of Yelling Lake and may represent either a deformed volcanic breccia or conglomerate. At Dadson Lake, bedded and massive felsic tuffs are interlayered with schistose mafic volcanic rocks and thin units of felsic tuff are interbedded with slates in a narrow supracrustal sliver 5 km northwest of Dadson Lake.

Intermediate tuffs 20 km east of Yelling Lake and 5 km northwest of Dadson Lake have identical ages of 2718 Ma (Nos. 31 and 42 of Table 3). No geochemistry is available for these rocks.

Oxford Lake Sedimentary Assemblage

Sedimentary sequences are exposed within a thin supracrustal sliver 5 km northwest of Dadson Lake and are also interpreted to mantle an oval sanukitoid pluton at Umisko Lake, Manitoba (*see* map in back pocket). Fine-grained slates are exposed in rare outcrops at the locality 5 km northwest of Dadson Lake. This sedimentary unit has a strong aeromagnetic expression (ODM-GSC 1967a), which is interpreted to indicate that iron formation is associated with the fine-grained clastic sedimentary rocks. Likewise, a strong magnetic anomaly at Umisko Lake (ODM-GSC 1967b) is centered over a pluton of the sanukitoid suite. Although sanukitoid plutons tend to have magnetic anomalies, the intensity of the anomaly at Umisko Lake suggests that sedimentary rocks including iron formation may occur at margins of the pluton.

Tuff interbedded with the slates northwest of Dadson Lake are 2718 Ma old (discussed above) and indicate a late depositional age for the sedimentary sequence. On this basis, the fine-grained clastic and chemical sedimentary sequences appear to be associated with the Oxford Lake sedimentary assemblage.

PLUTONIC ROCKS

Felsic plutonic rocks of the northern Superior area are subdivided into six major suites based on mineralogy (e.g., biotite tonalite, biotite granite and hornblende tonalite suites), composition (e.g., peraluminous and sanukitoid suites) and structure (gneissic suite). This subdivision follows the classification of Stone (1998) for plutonic rocks of the Berens River area of northwest Ontario. The various plutonic suites are listed in Table 5 together with characteristics of each suite including rock type, colour, grain-size, fabric, form, and inclusion types that are used to identify the suites in the field. Also shown in Table 5 are average mineral assemblages and known age determinations for each suite.

Typically, several subdivisions of each suite are made in the field on the basis of variations in mineralogy (e.g., biotite tonalite that is rich in potassium feldspar and gradational to biotite granite), mafic mineral content (e.g., mafic biotite tonalite) and texture (e.g., weakly gneissic biotite tonalite). These variations are listed in the geologic legend in MRD 135 and are shown by separate codes (12r, 12m and 12g, respectively) on the 1:50 000 scale geologic maps referenced in Figure 2. On the 1:250 000 scale map accompanying this report, the various subdivisions of each suite are grouped into the six major suites represented by map units 11 to 16.

The felsic plutonic suites are described individually below. Also included with plutonic rocks are descriptions of Proterozoic carbonatite intrusion and gabbro dikes.

Biotite Tonalite Suite

The biotite tonalite suite occurs widely, underlying 34% of the northern Superior area. Although several mineralogical and textural varieties are mapped, rocks of the biotite tonalite suite are typically medium grained, grey and foliated and composed of feldspars and quartz with less than 10% biotite (Table 5). The proportion of potassium feldspar varies significantly so that rock compositions scatter through the fields of tonalite and granodiorite and rarely granite on the ternary diagram of Streckeisen (1976; Figure 12a). A few mafic phases of the biotite tonalite suite are quartz diorite and quartz monzodiorite in composition.

Biotite tonalite occurs as large irregular to belt-like masses flanked by greenstone belts and intruded by younger plutonic rocks such as those of the hornblende tonalite and biotite granite suite (see map in back pocket). Biotite tonalite is gradational to tonalite gneiss, through development of a layered fabric with lithologic variation between layers. In places, rocks of the biotite tonalite suite are gradational to rocks of the biotite granite suite due to progressive increase in the proportion of potassium feldspar. Elsewhere, biotite granite crosscuts biotite tonalite. Contact relations of the hornblende tonalite and sanukitoid suites to the biotite tonalite suite are rarely observed, however rocks of the former suites crosscut biotite tonalite in a few localities where contacts are exposed. Likewise the contacts between biotite tonalite and greenstone sequences are rarely exposed; at Ponask Lake, conglomerate of the Sachigo assemblage is interpreted to unconformably overlie biotite tonalite. Clasts of biotite tonalite are observed in conglomerate of the Cross Lake assemblage.

Table 5. Characteristics of plutonic suites, northern Superior area.

Suite/Map Unit No.	Rock type	Colour	Grain Size	Fabric	Form and Occurrence/% of area	Inclusion type	Mineral Assemblage (average mineral proportion) [no. of samples]	Age (Ma)*
Biotite tonalite/12	biotite tonalite to granodiorite	white to grey	fine to coarse	Foliated to gneissic; quartz and feldspar megacrystic	Irregular to crescentic and lobate bodies; scattered/34	amphibolite, supracrustal xenoliths	Pl(56.2)+Qtz(26.4)+Bt(9.1)+Kfeld(8.3)+Mag(<1)+Ttn(<1)+Ep(<1)+Ap(<1)+Aln(<1)+Ilm(<1)+Zrn(<1) [76]	2863±0.7, 2855±5, 2822.0±0.8, 2733.7±1.7, 2722.5±2
Gneissic/11	biotite+hornblende tonalite to granodiorite	dark grey to white	variable between layers	Foliated; layered; folded	Belts, masses; scattered near greenstone/3	supracrustal xenoliths	Pl(49)+Qtz(25)+{Hbl+Bt}(13)+Kfeld(13)+Mag(<1)+Ep(<1)+Ttn(<1)+Ap(<1)+Aln(<1)+Ilm(<1)+Zrn(<1) [5]	2848±7, 2814±4, 2701.1+.4/-1.8
Hornblende tonalite/16	hornblende+biotite tonalite to granite	grey to white and pink	coarse	Foliated; granular; feldspar megacrystic	Irregular to elongate bodies of variable size; scattered/14	lensoid dioritic inclusions	Pl(54)+{Bt+Hbl}(19)+Qtz(19)+Kfeld(7)+Mag(1)+Ep(<1)+Ttn(<1)+Ap(<1)+Aln(<1)+Ilm(<1)+Zrn(<1) [33; tonalite]	2718.7±1.5, 2716.2±1.2, 2715±8
Biotite granite/15	biotite granodiorite to granite	white to pink	medium to coarse	Massive to weak magmatic layering	Dikes, irregular masses, oval batholiths; scattered/30	biotite tonalite, amphibolite	Pl(42)+Qtz(27)+Kfeld(25)+Bt(6)+Ttn(<1)+Mag(<1)+Ep(<1)+Ap(<1)+Ilm(<1)+Aln(<1)+Zrn(<1) [60]	2846±5, 2723±6, 2690 to 2711
Peraluminous (S-type)/13	biotite+muscovite granodiorite to granite	white	coarse to pegmatitic	Massive; mylonitic	Elongate to irregular bodies/<1	sediment	Pl(44)+K feld(24)+Qtz(29)+{Bt+Ms}(3)+Gt(<1)+Ap(<1)+Sil(<1)+Crd(<1)+Mnz(<1) [3]	
Sanukitoid/14	biotite+hornblende+pyroxene quartz diorite, tonalite, quartz monzodiorite, granodiorite, quartz monzonite, quartz syenite, granite	grey to pink and red	medium to coarse	Massive to weak magmatic layering	Oval plutons/4.5	hornblendite, amphibolite	Pl(43)+{Bt+Hbl+Cpx}(32)+Kfeld(20)+Qtz(2)+Mag(2)+Ttn(1)+Ap(<1)+Ep(<1)+Ilm(<1)+Py(<1)+Zrn(<1) [12; monzodiorite]	2714±8, 2710±6

Mineral abbreviations are from Kretz (1983); K feld=k feldspar

* sources of age determinations are references accompanying Table 3

Biotite tonalite varies widely in age from 2863 to 2722 Ma (Table 5). Together with the gneissic suite, some intrusions of the biotite tonalite suite appear to represent the oldest plutonic rocks in all terranes of the northern Superior area. Other intrusions of the biotite tonalite suite were emplaced in the Neoarchean (2733 to 2722 Ma) at the onset of widespread magmatism associated with the hornblende tonalite and biotite granite suites.

Geochemically, the biotite tonalite suite is calc-alkalic (Figure 13a), mildly metaluminous to peraluminous (ACNK[5] = 0.9 to 1.1) and subalkalic and mainly of the medium-K type of LeMaitre (1989; Figure 14a). In comparison with other plutonic suites, the biotite tonalite suite has a low Mg#[6] (.46) and is high in Al_2O_3 and Na_2O and low in LILE and REE (Table 6).

Rocks of the biotite tonalite suite are subdivided by terrane, as shown by symbols on geochemical diagrams, although only minor geochemical variations are recognized between samples from different terranes. Primitive-mantle-normalized multi-element profiles are strongly sloped from left to right (Figures 15a, b, and c) reflecting enrichment in the incompatible elements. Samples have trace element profiles that tend to be parallel to one another for the more compatible elements (*see* right sides of Figures 15a, b, and c) but cross each other for the more incompatible elements (*see* left sides of Figures 15a, b, and c). This characteristic may indicate that incompatible elements have been mobilized by alteration.

The multi-element profiles for biotite tonalite from all terranes have deep troughs for Nb and Ti, characteristic of plutonic rocks developed in magmatic arcs. Biotite tonalite from the Island Lake terrane is somewhat elevated in LILE and LREE compared to samples from other terranes and has variably developed troughs for Ba and Sr and possibly Eu. The depletion in Ba and Sr can be explained by fractionation of feldspar, which has positive distribution coefficients for these elements in a siliceous melt (Rollinson 1996).

With minor exceptions, rocks of the biotite tonalite suite have >14.5 wt% Al_2O_3 and < 1.5 ppm Yb (Figure 16) and are classified as high Al_2O_3, HREE-depleted (continental) tonalites (Arth 1979). Arth and Hanson (1972) and Arth and Barker (1976) proposed on the basis of trace element modeling that tonalitic magmas of the continental type can be produced by 20% melting at mid to lower crustal depths of basalt or amphibolite leaving an eclogitic or hornblende-rich residue. Subsequent work of Drummond and Defant (1990) indicated that dehydration melting of subducted oceanic crust within continental magmatic arcs is a likely environment for production of high-Al tonalitic magmas.

High-Al rocks of the biotite tonalite suite plot within the field of volcanic arc "granite" on the Y+Nb-Rb diagram of Pearce et al. (1984; Figure 17a). This, in addition to strong Nb and Ti troughs on multi-element profiles support the interpretation that the biotite tonalite suite was developed in continental magmatic arcs. The arcs were active at 2.86, 2.82 and 2.73 to 2.72 Ga as well as possibly earlier times in the northern Superior area (Table 5).

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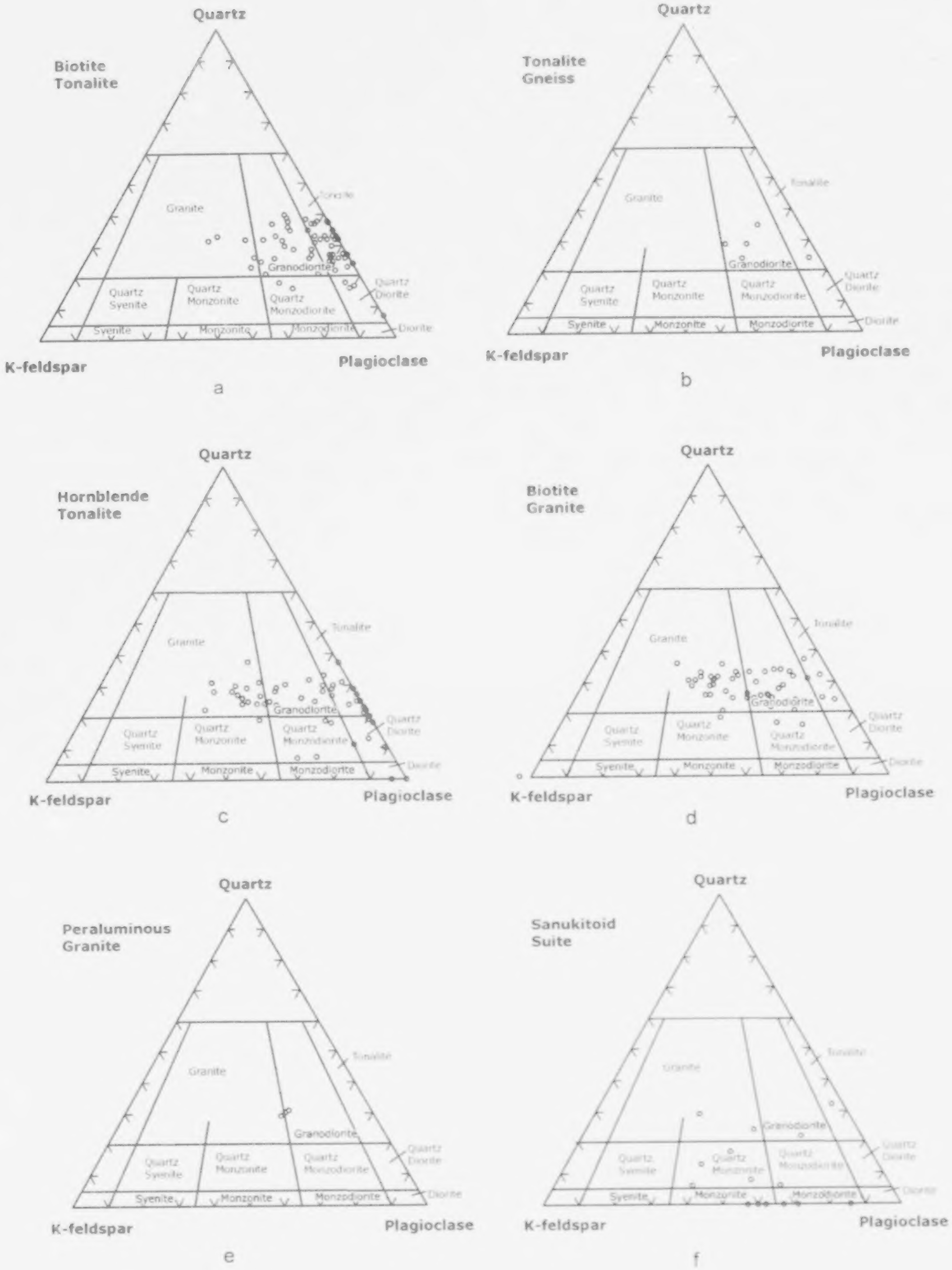
[5] ACNK=(wt% Al_2O_3/molecular weight Al_2O_3)/((wt %CaO/molecular weight CaO)+(wt %Na_2O/molecular weight Na_2O)+(wt % K_2O/molecular weight K_2O))

[6] Mg#=(MgO[weight %]/molecular weight MgO)/((MgO[weight %]/molecular weight MgO)+(FeO[weight %]/molecular weight FeO))



a b c d e f

Figure 12. Modal proportions of quartz, potassium feldspar and plagioclase for plutonic suites.

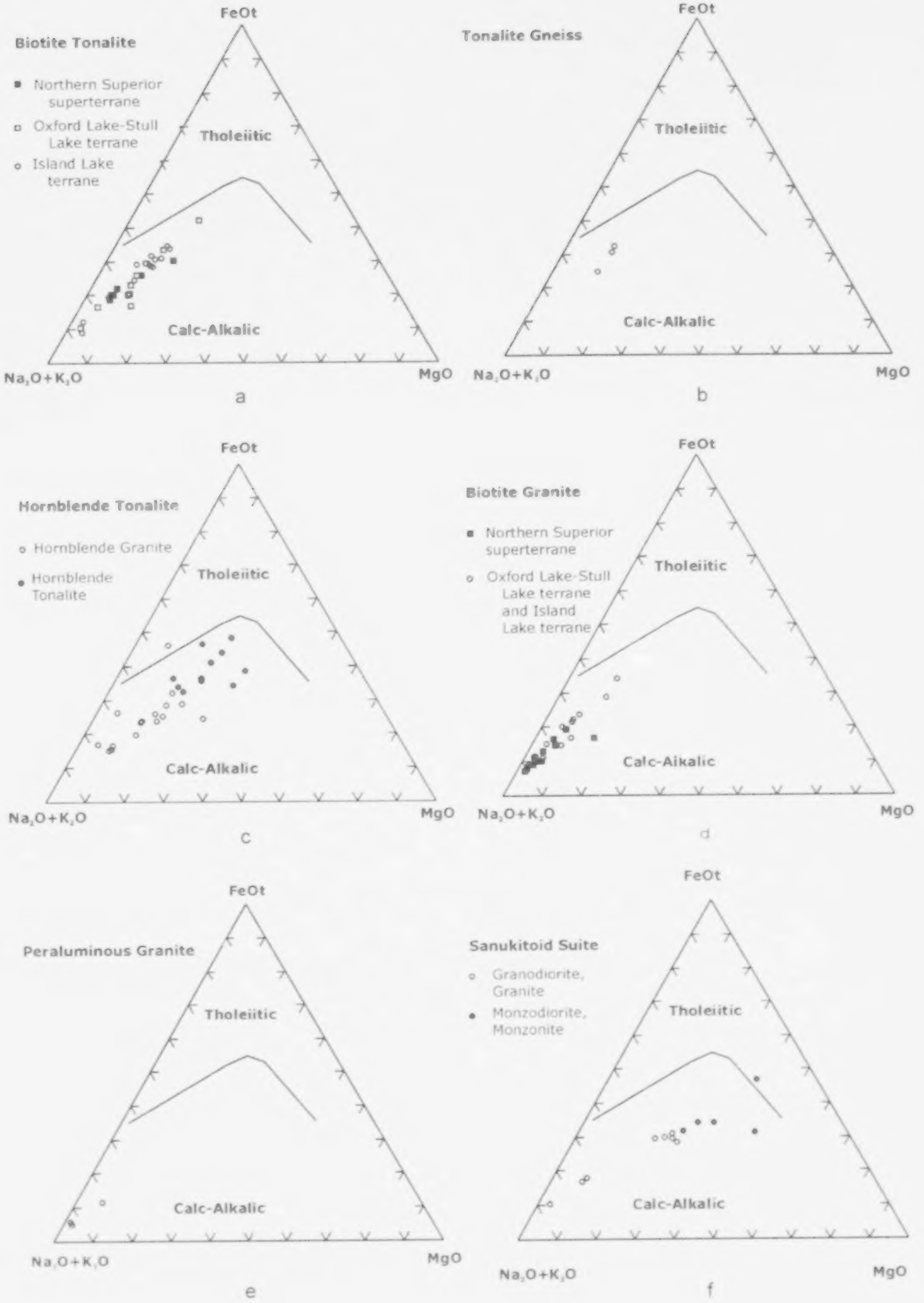


Figure 13. AFM plot for plutonic suites.

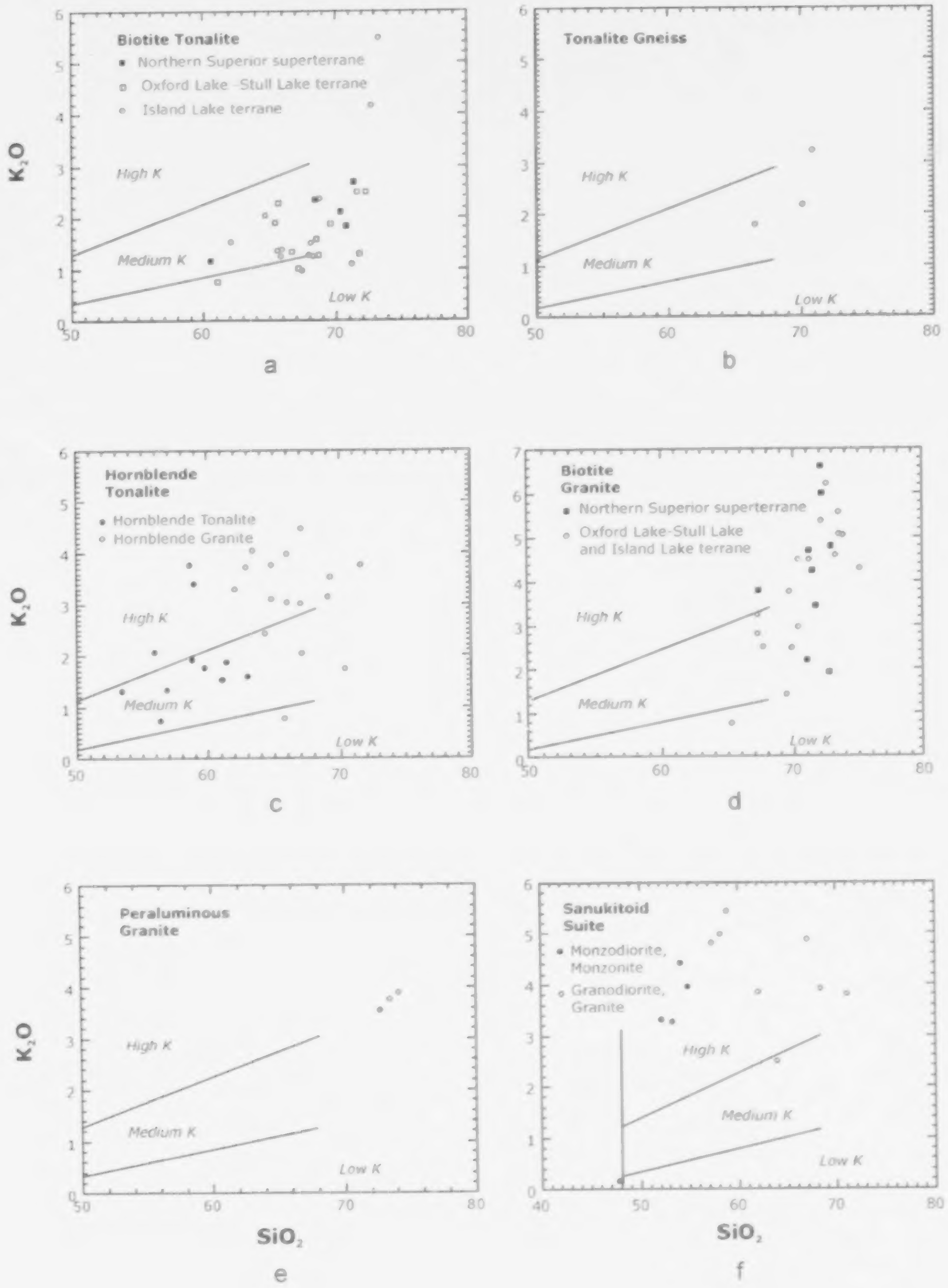


Figure 14. K$_2$O-SiO$_2$ plots for plutonic suites.

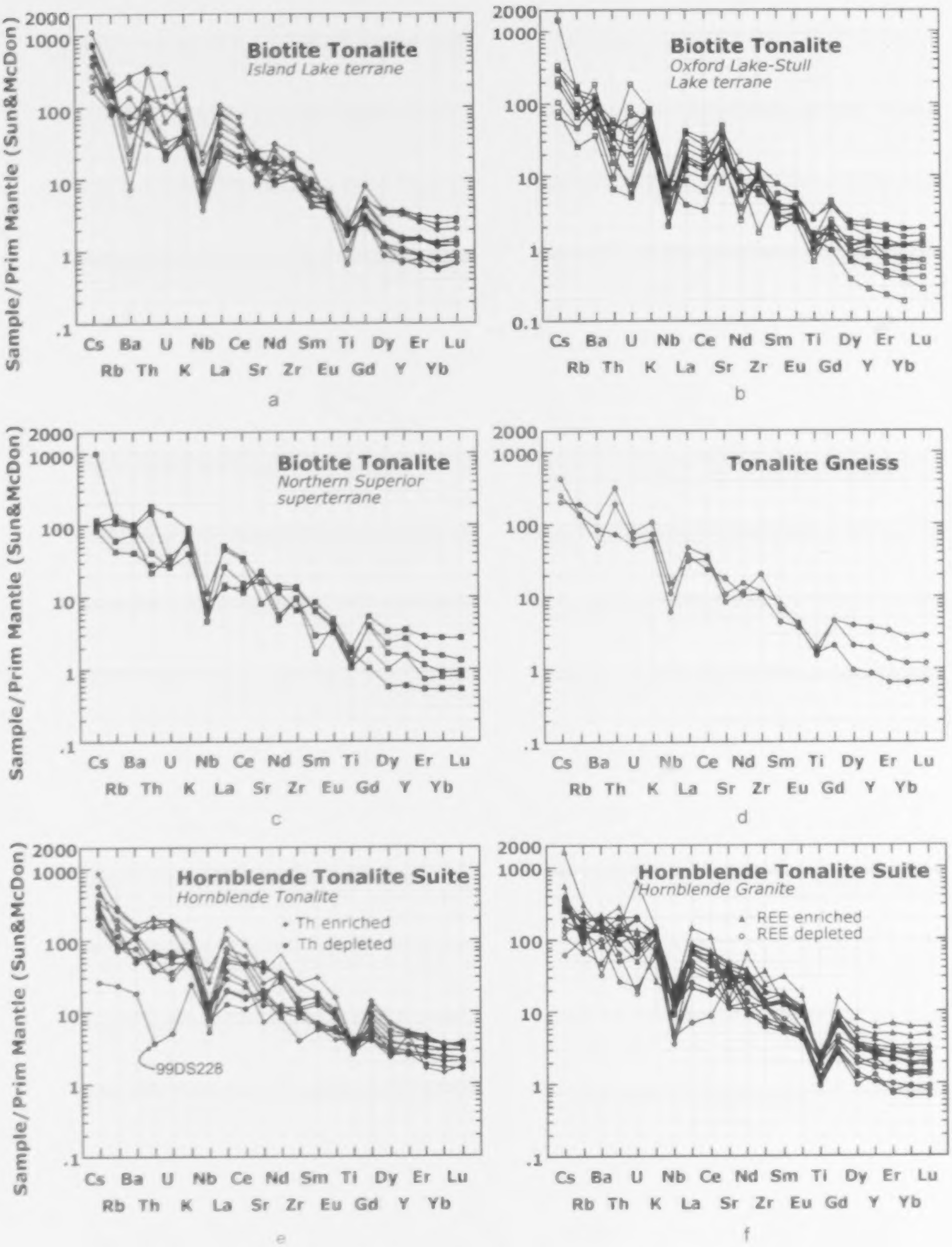


Figure 15. Multi-element plots for plutonic suites.

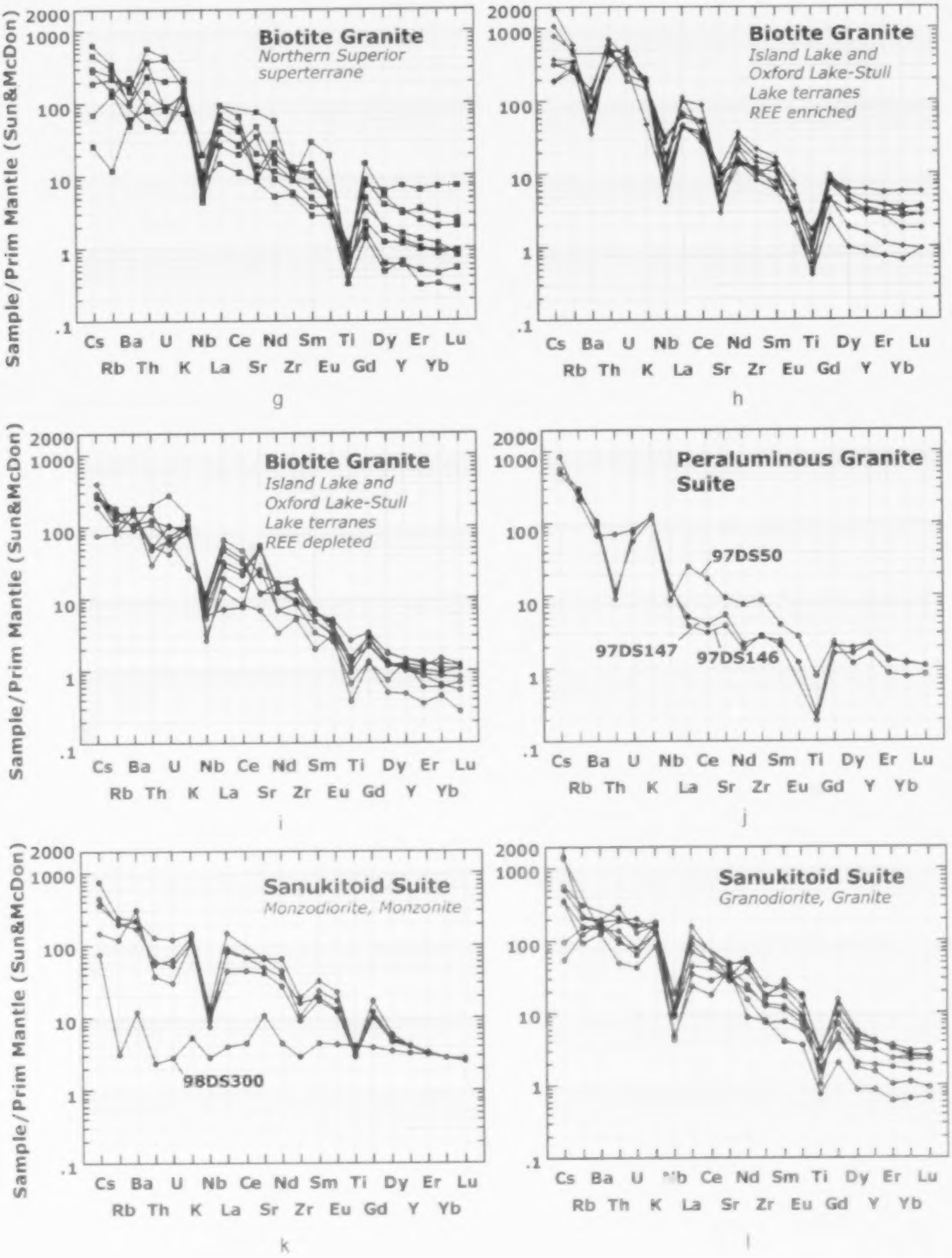


Figure 15 (continued). Multi-element plots for plutonic suites.

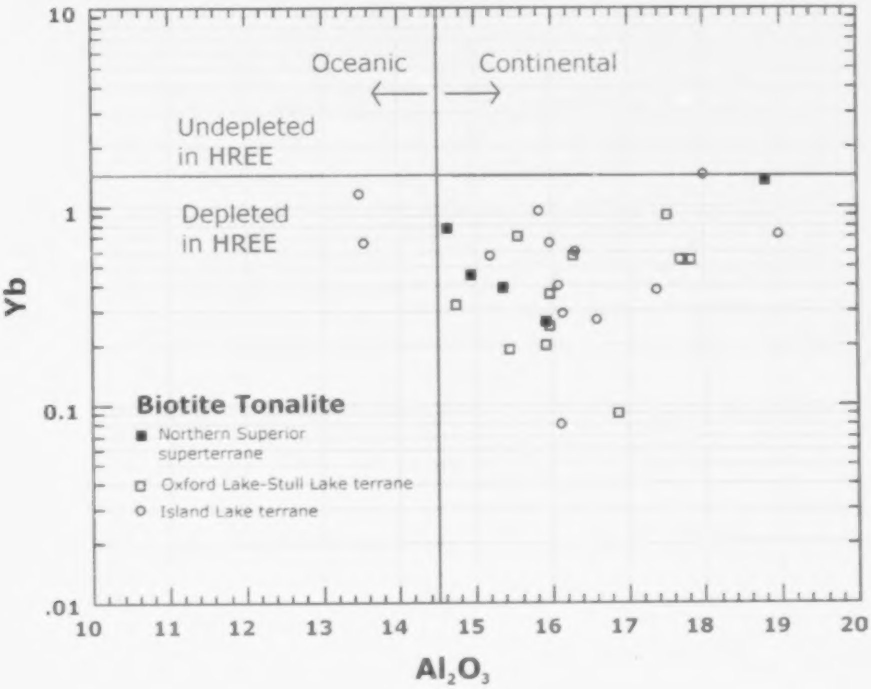


Figure 16. Yb-Al_2O_3 plot for the biotite tonalite suite.

Gneissic Suite

Although units of tonalite gneiss occur widely, they are rare and comprise only 3% of the northern Superior area. Tonalite gneiss is typically a heterogeneous layered and variably folded rock made up of several compositional and textural varieties of tonalite, diorite and amphibolite. Xenoliths of amphibolite representative of both deformed gabbro dikes and mafic lava flows are common in the gneisses. Felsic dikes of biotite tonalite and biotite granite cut tonalite gneiss.

Tonalite gneiss is spatially associated with biotite tonalite. Locally, gneisses occur adjacent to greenstone belts (e.g., southern Stull Lake: *see* map in back pocket) or else within belt-like domains possibly representing strained margins of eroded greenstone sequences (e.g., Yelling Lake area: *see* map in back pocket). Tonalite gneiss ranges compositionally from felsic varieties comprising mainly biotite tonalite and granodiorite (Figure 12b) to mafic varieties that have a large proportion of amphibolite. At the felsic end of the compositional spectrum, tonalite gneiss is gradational to biotite tonalite. At the mafic end of the compositional spectrum, tonalite gneiss is gradational to supracrustal rocks through increased proportion of xenoliths and development of continuous compositional layering representative of metamorphosed primary structures such as sedimentary bedding and lava flows.

Table 6. Average chemical analyses of plutonic suites and the Oxford Lake volcanic assemblage, northern Superior area.

Suite	11	12	13	14a,b	14c-e	15	16a, b	16c	9v
No. Analyses	3	29	3	5	8	26	11	16	29
Area	North Superior	North Superior	North Superior	North Superior	North Superior	North Superior	North Superior	North Superior	North Superior
oxide (wt%)									
SiO2	69.14	60.53	73.46	52.47	63.34	70.95	58.56	66.35	57.44
TiO2	0.37	0.33	0.09	0.73	0.43	0.19	0.73	0.38	0.67
Al2O3	15	16.18	14.66	14.33	14.61	15.02	16.18	15.25	16.5
Fe2O3	0.92	1.02	0.27	2.68	1.59	0.84	2.16	1.31	1.79
FeO	2.34	1.66	0.41	5.56	2.43	0.8	4.23	1.9	4.19
FeOT	3.16	2.6	0.66	7.96	3.86	1.56	6.18	3.08	5.8
Fe2O3T	3.52	2.89	0.73	8.85	4.29	1.73	6.86	3.42	6.45
MnO	0.03	0.03	0.03	0.12	0.05	0.02	0.1	0.06	0.11
MgO	1.32	1.05	0.31	7.81	2.92	0.58	3.57	1.75	2.99
CaO	2.85	3.39	1.15	7.34	3.68	1.87	5.71	3.1	5.29
Na2O	3.9	4.71	4.26	3.45	4.07	4.14	4.15	4.47	3.49
K2O	2.39	1.87	3.73	3.03	4.28	3.96	1.94	3.11	2.28
P2O5	0.07	0.07	0.02	0.42	0.25	0.05	0.23	0.13	0.29
H2Op	0.49	0.51	0.41	1.27	0.79	0.4	1.12	0.58	1.76
H2Om	0.15	0.17	0.15	0.23	0.17	0.18	0.19	0.21	0
CO2	0.18	0.18	0.18	0.21	0.38	0.19	0.36	0.15	2.5
S	0.02	0.02					0.02	0.01	
LOI	0.67	0.72	0.71	1.02	1.12	0.77	1.19	0.92	4.22
Total	99.25	99.24	99.15	99.57	99.05	99.28	99.2	98.94	99.74
ACNK	1.06	1.02	1.12	0.64	0.83	1.04	0.84	0.94	0.95
ANK	1.66	1.67	1.34	1.7	1.3	1.37	1.86	1.44	2.17
Mg	0.46	0.46	0.49	0.67	0.59	0.45	0.54	0.54	0.5
Ti (ppm)	2257.03	2016.99	527.55	4454.87	2657.89	1180.79	4497.65	2356.15	4216.38
P (ppm)	295.59	331.98	73.76	1856.99	1136.39	206.47	1008.19	576.83	1348.34
K (ppm)	20206.75	15742.67	31497	25759.98	36349.15	33348.64	16493.24	26428.5	20038.14
Ba (ppm)	613.34	622.42	636.67	1319.4	860.88	705.99	680.09	998.75	747.41
Rb	99	72	165	98	102	141	88	97	71
Sr	262	439	133	1002	790	437	502	618	672
Cs	2.3	2.9	5	2.9	4.8	3.1	2.5	2.7	4.1
Be				0.6	0.9		0.4		0.2
Li	42.3	38.2	53	37	33.8	38.7	41.1	34.1	22.7
Mo		4.65		2	4.75	1.96	0.91	5.22	7.17
Ta	0.6	0.4	1.5	0.3	0.5	0.8	0.9	0.8	0.7
Hf	3.3	3.5		3.1	5.3	4.1	3.8	4.5	4.6
Ga	18.5	19.8	0	17.4	19.1	17	21	20.2	19.9
Nb	7.8	5.2	8.6	6.2	7.3	7.1	9.7	8	9.2
Sn		0.93					1.13		0.21
Zr	163	132	53	129	163	123	151	159	169
Co	7	5.6		36.6	13.4	0.9	20.5	7.6	18.7
Cr	28	26	8	259	133	17	89	59	88
Cu	19	11.8	39.3	53.2	24.3	1.4	30.1	8.4	42.6
Ni	16	12.5	0	167.2	44.8	5.2	53.9	27.5	61.4
V	40	39	5	167	71	17	119	47	96
Zn	53.7	59.6	21	93.8	64.5	38.8	102.5	61.8	96.9
Pb	15.8	13.8	14	14.2	20.7	24.6	13.1	18	14.9
Th	16.7	7.9	2.6	4.9	13.7	23.6	8.3	11	10.6
U	1.35	1.22	1.51	1.14	2.43	4.48	1.9	2.86	2.36
Sc	4.3	3.9	0.7	18.6	8.6	1.9	12.8	6.1	10.2
Y	10	7	9	18	13	10	17	12	17
La (ppm)	26.88	23.62	7.81	50.43	56.34	32.74	39.46	39.79	46.58
Ce	55.64	42.82	14.3	66.3	31.87	51.94	61.76	79.47	86.25
Pr	5.11	5.32	1.5	12.94	12.91	6.68	9.41	9.41	12.56
Nd	17.02	17.63	5.31	49.39	45.63	22.56	33.64	33.58	45.66
Sm	2.84	2.63	1.24	8.71	7.08	3.69	5.62	5.34	7.93
Eu	0.68	0.72	0.22	2.33	1.71	0.73	1.35	1.31	1.91
Gd	2.26	1.95	1.18	6.45	4.95	2.76	4.4	3.85	5.39
Tb	0.34	0.23	0	0.81	0.59	0.34	0.58	0.45	0.69
Dy	1.75	1.22	1.16	3.54	2.44	1.8	3.06	2.4	3.46
Ho	0.35	0.21	0.17	0.66	0.44	0.31	0.56	0.42	0.64
Er	0.87	0.61	0.55	1.57	1.09	0.89	1.53	1.19	1.77
Tm	0.12	0.08	0.08	0.21	0.15	0.12	0.21	0.16	0.23
Yb	0.73	0.55	0.52	1.39	1.02	0.86	1.35	1.12	1.53
Lu	0.12	0.08	0.07	0.2	0.15	0.13	0.2	0.17	0.25
La/Yb$_N$	36.66	41.63	13.35	25.93	40.32	36.60	20.38	28.15	22.75

The crystallization age of tonalite gneiss ranges from 2848 to 2701 Ma (Table 5). Samples of tonalite gneiss from north of the Kenyon fault contain inherited zircon grains with ages as old as 3572 Ma and 3292 Ma (Samples 38 and 40 of Table 3). These represent assimilated remnants of the Paleoarchean Northern Superior Superterrane.

Three samples of tonalite gneiss from the Northern Superior Superterrane were analyzed (*see* file *North Superior Geochemistry 2000* on MRD 135; unit 11) and show chemical characteristics broadly comparable to those of the biotite tonalite suite. Tonalite gneisses are calc-alkalic (Figure 13b) and of the medium-K type (Figure 14b) and have a low Mg# of 0.46 and are mildly peraluminous (ACNK=1.0 to 1.1). The gneisses have Al_2O_3>14.5 wt% and Yb<1.5 ppm and are classified as continental-type tonalite. Compositionally, the gneisses have higher FeOT, MgO and transition metals (V, Cr, Mn, Co, Cu, Ni, Zn) than biotite tonalite (*see* Table 6) reflecting a higher proportion of mafic minerals.

Multi-element profiles of tonalite gneiss are steeply sloped from left to right (Figure15d) with troughs at Nb and Ti, and are overall comparable to those of biotite tonalite. On the Rb-Y+Nb diagram of Pearce et al. (1984; Figure 17b), tonalite gneiss samples plot within the field of volcanic arc "granite". Tonalite gneiss magmas have apparently originated by a mechanism similar to that envisaged for the petrogenesis of biotite tonalite magmas, which involves partial melting of subducted oceanic crust in continental magmatic arcs. The bulk of tonalitic magma was emplaced within large plutonic masses that crystallized as massive to foliated biotite tonalite. Development of the layered fabric characteristic of tonalite gneisses has apparently occurred within zones of high strain typically associated with the contact areas of greenstone sequences or broad fault zones. The various phases of tonalitic magma that intruded at high temperature within the strain zones would have been tectonically mixed and stretched giving rise to a gneissic fabric with incorporation of variable proportions of supracrustal xenoliths.

Hornblende Tonalite Suite

Rocks of the hornblende tonalite suite comprise 14% of the northern Superior area and occur in all terranes, typically in the form of elongate to belt-like bodies. Rocks of the hornblende tonalite suite show a weak spatial association with greenstone belts and rarely occur as oval plutons within greenstone belts such as at northern Stull Lake. Rocks of the hornblende tonalite suite vary in composition through fields of tonalite, granodiorite and granite with some mafic varieties having the composition of quartz diorite and quartz monzodiorite (Figure 12c). Two major subdivisions of the hornblende tonalite suite are mapped in the field. These are hornblende tonalite and megacrystic hornblende granite, the latter of which is distinguished by large potassium feldspar megacrysts. Hornblende tonalite and hornblende granite are coarse-grained, granular, mesocratic rocks with distinct lensoid dioritic inclusions. Megacrystic hornblende granite has up to 20% more quartz and potassium feldspar with correspondingly less plagioclase and mafic minerals than hornblende tonalite.

In spite of wide compositional variation, the hornblende tonalite suite contains a relatively high proportion of mafic minerals (average of 19% amphibole and biotite for hornblende tonalite; Table 5) with biotite predominating over amphibole in rocks that are strongly foliated. The compositions of plagioclase and amphibole grains from 42 samples of the hornblende tonalite suite were determined by microprobe for application of the Al-in-hornblende barometer (*see* "Al-in-Hornblende Barometry"). The mineral compositions are included in the digital files *Feldspar* and *Amphibole* on MRD 135. Plagioclase and amphibole grains are mainly of oligoclase and magnesio-hornblende composition, respectively.

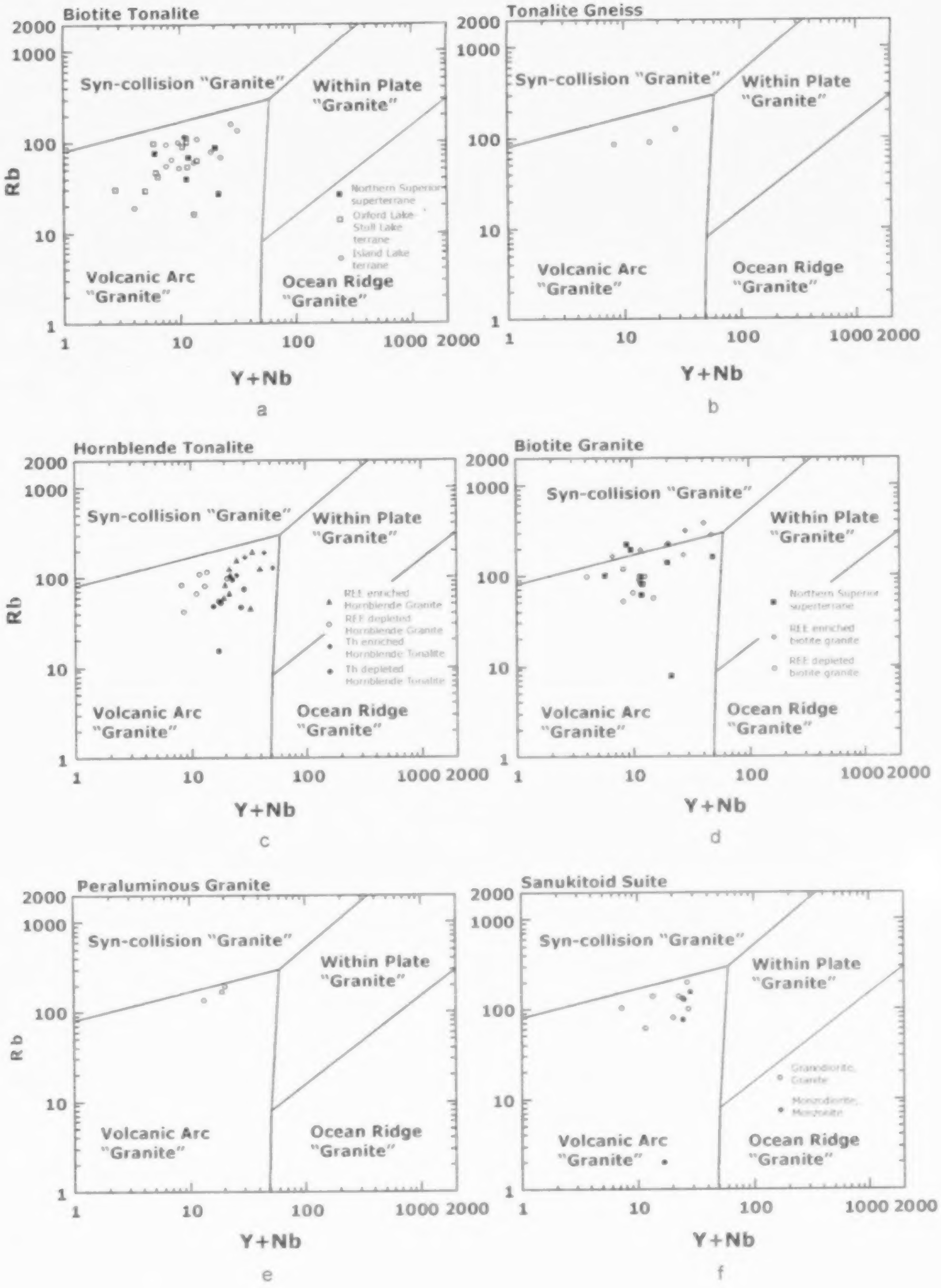


Figure 17. Rb-(Y+Nb) plot for plutonic suites.

Hornblende tonalite crosscuts biotite tonalite at Ellard Lake and is, in turn, cut by dikes of biotite granite. Boulders of hornblende tonalite occur in conglomerate of the Cross Lake assemblage at Stull Lake but are not observed in conglomerate of the Sachigo assemblage at Ponask Lake. Geochronologic studies of three intrusions indicate a narrow range of crystallization ages from 2719 to 2715 Ma (*see* Tables 3, 5) for the hornblende tonalite suite. The age determinations support field relations and imply that the majority of hornblende tonalite intrusions were emplaced at a late stage in the tectonic evolution of the area. At some localities however, strongly foliated hornblende tonalite is intermixed with 2.85 Ga biotite tonalite and possibly represents an earlier generation of the hornblende tonalite suite.

Rocks of the hornblende tonalite suite are calc-alkalic (Figure 13c), metaluminous (ACNK=0.7 to 1.05) and of medium- to high-K type (Figure 14c) with an average Mg# of 0.54 (Table 6). The hornblende tonalite suite shows considerable major-element variation and is divided into subgroups of tonalite to granodiorite and granodiorite to granite composition. These subgroups of the hornblende tonalite suite are shown by separate symbols in Figures 13 to 15 and by separate average analyses in Table 6. In comparison with biotite tonalite, hornblende tonalite (sensu stricto) is enriched in FeOT, MgO, CaO, transition metals and REEs and has lower concentrations of SiO_2 and Na_2O (Table 6). Hornblende granite is enriched in SiO_2, K_2O and LILE and depleted in FeOT, MgO, CaO and transition metals compared to hornblende tonalite.

Multi-element profiles for hornblende tonalite and hornblende granite of the hornblende tonalite suite are steeply sloped from left to right with deep troughs for Nb and Ti (Figures 15e, f). Rocks of hornblende tonalite to granodiorite composition are subdivided according to enrichment or depletion in Th and U (*see* Figure 15e). Included with the Th-depleted group is sample 99DS228 from a mafic phase of hornblende tonalite east of Stull Lake that is strongly depleted in Th, U and LILE. Hornblende granite samples are somewhat arbitrarily subdivided into groups that are REE-enriched and REE-depleted compared to an average composition of hornblende granite (*see* Figure 15f). Samples of hornblende granite have erratic concentrations of Cs, Rb, Ba, Th and U shown by crossed profiles in Figure 15f. The highly variable concentration of incompatible elements is interpreted to indicate that these elements may have been mobilized at one or more stages during the evolution of hornblende granite.

In view of the mobility of Th and U in hornblende granite, it is possible that the Th-enriched and Th-depleted varieties of hornblende tonalite have also developed due to alteration. On the other hand, the Th-enriched hornblende tonalite is also enriched in REEs. Th and REE enrichment are characteristics of crustal rocks. Possibly, the Th- and REE-enriched varieties of hornblende tonalite have developed due to assimilation of older crustal material such as biotite tonalite.

The hornblende tonalite suite is similar in age and chemical composition to the Oxford Lake volcanic assemblage. The right-hand column of Table 6 shows an average composition of 29 samples of the Oxford Lake volcanic assemblage in the Stull–Swan lakes greenstone belt. In comparison, the majority of major and trace element concentrations in the average volcanic rock fall intermediate between those of hornblende tonalite and hornblende granite. Exceptions are somewhat higher Al_2O_3, P_2O_5, Sr, Cs, Zr, Cu, Ni and REEs and lower Na_2O, Rb and Li in the average volcanic rock than in average hornblende tonalite or average hornblende granite. These chemical differences can be partly explained by fractionation of minor phases such as apatite, zircon, titanite and allanite in the magmas that crystallized as plutonic rocks and fractionation of hornblende and biotite in the magmas that erupted at surface (Zen 1986; Table 4.3 of Rollinson 1996).

Multi-element profiles of the various sub units of the Oxford Lake volcanic assemblage at Stull Lake correspond closely with those of the hornblende tonalite suite (compare Figures 10a to f with Figures 15e, f). The Gilleran Lake intrusion, which is closely associated with the Oxford Lake volcanic assemblage, is chemically similar to rocks of the hornblende tonalite suite that are most enriched in trace element. The chemical and age similarities to the arc-generated volcanic magmas of the Oxford Lake volcanic assemblage as well as Rb-Y+Nb systematics for hornblende tonalite (Figure 17c) indicate that the hornblende tonalite suite likely developed within magmatic arcs.

The development of medium- to high-K, LILE and LREE-enriched magmas with moderately high Mg# after about 2730 Ma probably reflects a major switch in the source of magmas as a result of terrane interactions. At an early stage, biotite tonalite magmas were generated by melting of subducted oceanic crust, however, the cessation of subduction after terrane abutment would have caused the melting regime to switch to the previously metasomatized mantle wedge. The transition from early hornblende tonalite to later hornblende granite of the hornblende tonalite suite may reflect an evolution in magma compositions as the source of melting switched from the subducted slab to the mantle wedge. In any case, the enriched magmas of the hornblende tonalite suite (hornblende granite phases), the Oxford Lake volcanic assemblage and subsequently the sanukitoid suite (discussed below) appear to have been drawn from an enriched source such as the mantle wedge. The widespread occurrence of plutons of the hornblende tonalite suite implies that magmas derived from the enriched source are more voluminous than would be estimated on the basis of the somewhat restricted Oxford Lake volcanic assemblage.

Biotite Granite Suite

The biotite granite suite occurs widely and underlies 30% of the northern Superior area. Intrusions of biotite granite are highly variable in size and shape. They include large, irregular to elongate batholiths such as between Pierce Lake and Little Sachigo Lake, oval plutons such as north of Blackbear Lake (*see* map in back pocket) and dikes and stocks cutting most other rock types at various localities.

Rocks of the biotite granite suite are typically medium- to coarse-grained, white to pink and massive. Inclusions of biotite tonalite and tonalite gneiss are common in biotite granite. On average, the major mineral constituents of the biotite granite suite are plagioclase with somewhat lesser proportions of quartz and K feldspar and a few percent biotite (*see* Table 5). The compositions of rocks assigned to the biotite granite suite are mainly granodiorite and granite with a few samples of tonalite, quartz monzodiorite and quartz monzonite (Figure 12d).

The crystallization ages of three samples of the biotite granite suite range from 2846 to as young as 2690 Ma (*see* Tables 3 and 5). At most localities in the northern Superior area, rocks of the biotite granite suite crosscut other plutonic suites which implies that the majority of biotite granite intrusions were emplaced in the late Neoarchean (2.72 to 2.69 Ga). These crosscutting relations are supported by geochronologic studies in other parts of the western Superior Province such as the Berens River area where ages for the biotite granite suite are younger than those for most other suites with the exception of the sanukitoid suite (Corfu and Stone 1998). In this context, the 2846 Ma biotite granite batholith in the Northern Superior Superterrane is unusual and provides an indication that biotite granite magmas were generated at an earlier time in the Northern Superior Superterrane than in most other parts of the western Superior Province.

Geochemically, rocks of the biotite granite suite are calc-alkalic due to strong enrichment in Na_2O and K_2O (Figure 13d) and are variable from the medium-K to high-K types (Figure 14d). They range from mildly metaluminous to mildly peraluminous (ACNK=0.9 to 1.1) and have a fairly low Mg# of .45 (Table 6). On average, biotite granite is enriched in SiO_2, K_2O, Pb, Th and U and depleted in FeOT, MgO, CaO and transition metals compared to most other plutonic rocks.

For purposes of analysis and display, samples of biotite granite from the Northern Superior Superterrane are separated from those of the Oxford Lake–Stull Lake and Island Lake terranes. Biotite granite from the latter terranes is further subdivided according to enrichment or depletion in REE. Where normalized to primitive mantle (Figures 15g to i), biotite granite samples show multi-element profiles similar to most other plutonic rocks. These are steeply sloped from left to right with deep troughs for Nb and Ti. REE-enriched biotite granite has troughs for Ba and Sr in addition to those for Nb and Ti. Rare earth element profiles for the REE-enriched biotite granite have negative Eu anomalies.

Several generalizations can be made regarding the petrogenesis of biotite granite from the primitive-mantle-normalized multi-element profiles. Firstly, depletion in Nb and Ti indicates that the magmas were generated in magmatic arcs or were derived by melting of material that had been previously formed in magmatic arcs. The biotite granite has elevated Rb causing samples to plot partly within the fields of volcanic arc "granite" and syn-collisional "granite" (Figure 17d). This and the generally peraluminous composition of biotite granite are more similar to the crustally derived peraluminous granite suite than to arc-derived suites such as biotite tonalite and imply that the Nb and Ti depletion are characteristics inherited from the source. In this case, the source is likely to be biotite tonalite, which developed earlier in magmatic arcs and occurs widely as inclusions in biotite granite. Secondly, depletion in Ba, Sr and Eu and general REE enrichment implies that the REE-enriched biotite granite has fractionated feldspars or has separated from a feldspar-bearing residuum. Thirdly, biotite granite from the Northern Superior Superterrane is partly of the REE-enriched variety and partly of the REE-depleted variety and does not seem to be compositionally distinct from younger granite in other terranes. This implies that the processes of granite magma generation were not much different at 2.85 Ga in the Northern Superior Superterrane than at 2.7 Ga in the amalgamated Superior Province.

Peraluminous Suite

The peraluminous suite, alternately known as the S-type granite suite or two-mica granite suite is rare in the northern Superior area. Although dikes of peraluminous granite are noted within all greenstone belts, the principal occurrences are in the Ponask Lake greenstone belt at Pierce Lake and adjacent to sedimentary arms of the Stull–Swan lakes greenstone belt 10 km west of Meston Lake (*see* map in back pocket). The Pierce Lake occurrence may represent an eastern extension of peraluminous granite noted within the Ponask Lake greenstone belt at Red Sucker Lake, Manitoba (Chackowsky and Cerny 1984).

Peraluminous granite is coarse-grained to pegmatitic and is composed of sub-equal proportions of plagioclase, quartz and K-feldspar with a few percent of biotite and muscovite and accessory garnet, cordierite and apatite (*see* Table 5 and Figure 12e). Peraluminous granite bodies are spatially associated with sedimentary sequences and can be transitional to sedimentary rocks through contact zones marked by a high proportion of sedimentary inclusions and migmatite.

Peraluminous granite intrusions of the northern Superior area are undated and can be variable in age. The peraluminous granite west of Meston Lake is inferred to be somewhat younger than the 2.72 Ga clastic sedimentary sequences of the Oxford Lake sedimentary assemblage with which it is associated. In contrast, peraluminous granite in the Ponask Lake greenstone belt is associated with sedimentary sequences of the Pierce assemblage and can be as old as 2.86 Ga (*see* Table 3).

Three samples of the peraluminous suite are calc-alkalic (Figure 13e), of the high-K type (Figure 14e) and are moderately peraluminous (ACNK=1.0 to 1.2) with a low Mg# of 0.49. Peraluminous granite is strongly enriched in SiO_2, K_2O, Rb and Li and is depleted in FeO, MgO, CaO, Sr, Zr, transition metals and REE compared to most other plutonic suites (*see* Table 6).

Multi-element profiles for the peraluminous granite suite are steeply sloped from left to right (Figure 15j) although HREE tend to be flat at approximately primitive mantle values. The multi-element profiles have troughs for Ti and the sample from Pierce Lake (97DS50) has a trough at Nb. The two samples from west of Meston Lake (97DS146 and 97DS147) are depleted in Th and LREE but are enriched in HREE compared to the Pierce Lake sample.

Sample 97DS50 is representative of the SP3-type peraluminous granite and samples 97DS146 and 97DS147 are representative of the SP4-type peraluminous granite of Sylvester (1994). SP3-type peraluminous granite has higher concentrations of CaO, TiO_2, FeOT, MgO, Sr, Ba, LREE and lower mean concentrations of Na_2O, Rb, Y, Nb, Cs, HREE and Ta than SP4-type peraluminous granite. Low levels of CaO, Sr and Na_2O in peraluminous granite compared to other plutonic suites is interpreted as chemical evidence that peraluminous melts are derived from sedimentary protoliths, which are depleted in these elements (Chappel and White 1974). Sylvester (1994) regarded SP3-type granite as having originated by partial melting of greywacke-pelite sources leaving a garnet-bearing residuum to account for HREE depletion. Higher Na_2O and lower CaO and Sr in SP4-type melts are attributed to the involvement of other material such as felsic volcanic rocks in the source for these magmas.

Elevated levels of Rb cause samples of peraluminous granite to plot in the transitional area between volcanic arc and syn-collision "granite" in the tectonic discrimination diagram of Pearce et al. (1984; Figure 17e).

Sanukitoid Suite

Rocks of the sanukitoid suite occur within distinct oval plutons associated with the Ellard Lake and Yelling Lake greenstone belts and collectively comprise 4.5% of the northern Superior area. Plutons of the sanukitoid suite are distinguished by high aeromagnetic anomalies (ODM-GSC 1967a, b) and are aligned east-southeast such as between Echoing Lake and Foster Lake and between Umisko Lake and Dadson Lake (*see* map in back pocket).

Rocks of the sanukitoid suite vary substantially in composition and include quartz-undersaturated phases of monzodiorite and monzonite as well as quartz-saturated phases of quartz monzodiorite, quartz monzonite, tonalite, granodiorite and granite (Figure 12f). The quartz-undersaturated phases have up to 40% mafic minerals comprising clinopyroxene+ hornblende+biotite. In contrast, quartz-rich phases such as granodiorite and granite are devoid of clinopyroxene and contain up to 10% hornblende+biotite. Typically, the quartz-undersaturated phases represent the oldest component of sanukitoid magmas and occur at margins of the plutons giving way to quartz-saturated phases in pluton interiors. Rocks of the sanukitoid suite are medium- to coarse-grained, pink to red and massive. Dark inclusions of gabbro and hornblendite are characteristic of the sanukitoid suite.

Sanukitoid plutons crosscut most other rock types. Intrusions of the sanukitoid suite east of Yelling Lake and at Echoing Lake have ages of 2714 and 2710 Ma (Nos. 33 and 21 of Table 3) and indicate that sanukitoid magmas were emplaced broadly contemporaneous with the late alkalic stages of volcanism of the Oxford Lake volcanic assemblage.

Samples of the sanukitoid suite show considerable compositional variation and are subdivided into quartz-undersaturated and quartz-saturated phases for purposes of analysis and display. Quartz-undersaturated phases are depleted in alkalis compared to quartz-saturated phases and one sample of mafic monzodiorite plots within the tholeiitic field (Figure 13f). With two exceptions, samples of the sanukitoid suite are of the high-K type (Figure 14f). Rocks of the sanukitoid suite are metaluminous (ACNK=0.6 to 1.0) with higher ACNK values common to quartz-bearing rocks. Monzodiorite and monzonite have a high average Mg# (0.67) decreasing to 0.59 in granodiorite and granite (*see* rock types 14a,b and 14c-e, respectively, in Table 6).

Rocks of the sanukitoid suite are enriched in FeOT, MgO, P_2O_5, LILE, transition metals and REEs compared to most other plutonic suites (*see* Table 6). Multi-element profiles are steeply sloped from left to right with deep troughs for Nb and Ti (Figures 15k, l). Sample 98DS300 of a mafic phase of the sanukitoid pluton at Ellard Lake has a flat trace element profile (*see* Figure 15k) and may represent a primitive component of sanukitoid magmas.

Shirey and Hanson (1984) interpreted the high FeOT, MgO, and transition metals as evidence that quartz-undersaturated phases of sanukitoid magmas originated by partial melting of a peridotitic mantle source. These authors explained the high LILE and REE abundances in the melts as evidence that the mantle sources had been metasomatically enriched in these elements prior to melt extraction. Further petrogenetic modeling by Stern et al. (1989) and Stern and Hanson (1991) confirmed the interpretation of Shirey and Hanson (1984) and proposed that the quartz-saturated phases of sanukitoid magmas have originated by fractionation of the quartz-undersaturated phases. Stevenson, Henry and Gariépy (1999) found depleted mantle-like Nd isotopic compositions in quartz-undersaturated phases and enriched Nd values in quartz-saturated phases of sanukitoid plutons in the western Superior Province. They concluded that the quartz-saturated phases had evolved by assimilation of crustal material as well as through fractionation from monzodioritic parent material.

Monzodiorite within the sanukitoid pluton east of Yelling Lake has ε_{ND} values of -0.55 and –1.71 whereas granodioritic parts of the sanukitoid pluton at Echoing Lake have ε_{ND} values of 0.92 and 1.15 (Skulski et al. 2000). These data are interpreted to indicate that the pluton east of Yelling Lake has assimilated older crustal material of the Northern Superior Superterrane whereas magmas of the Echoing Lake intrusion did not interact with older crust during emplacement at 2710 Ma. Evidently, crustal contamination has occurred in the quartz-undersaturated (monzodioritic) phases of the intrusion east of Yelling Lake.

Sanukitoid plutons plot within the field of volcanic arc "granites" on the diagram of Pearce et al. (1984; Figure 17f). The intrusion east of Yelling Lake must have been emplaced within a continental magmatic arc at the margin of the Northern Superior Superterrane where crustal assimilation was possible. In contrast, the intrusion at Echoing Lake could have intruded into an oceanic arc prior to accretion of the Oxford Lake–Stull Lake terrane to the Island Lake terrane.

Carbonatite

An oval intrusion (approximately 5 km diameter) of carbonatite occurs north-centrally in the northern Superior area (*see* map in back pocket). The carbonatite intrusion is completely covered with surficial materials but is distinguished by a strong aeromagnetic anomaly (ODM-GSC 1967a, b). The carbonatite intrusion (informally know as the "Carb" Lake carbonatite) appears to have been emplaced into biotite tonalite at or near to the South Kenyon fault although poor exposure precludes accurate definition of contact relations.

Most of what is known about the Carb Lake carbonatite is derived from exploration work by Big Nama Creek Mines Limited who completed a magnetometer survey and drilled 4 diamond-drill holes totaling 564 m in 1968 (Assessment Files, Resident Geologist's Office, Red Lake). R. Sage recovered the diamond-drill core and completed a geologic report including petrographic and geochemical data (Sage 1987) using samples derived from the diamond-drill core and boulders collected by Bennett and Riley (1969) and believed to be derived from the intrusion.

Sage (1987) described the diamond-drill core as layered and locally vuggy and composed of coarse white sövite (a carbonatite rock composed of 50% or more calcite) alternating with layers of silicocarbonatite (a carbonate-rich igneous rock composed of 50% or more silicate and oxide minerals) with local magnetite layers. The diamond-drill core recovered from the "Carb" Lake carbonatite is stored partly in the Drill Core Library at the Resident Geologist's Office in Kenora and partly with the Royal Ontario Museum in Toronto. Approximately 10 kg of material from Kenora was crushed and processed to identify heavy minerals (specific gravity greater than 3.1) associated with the carbonatite (*see* Stone 2001 for a description of methods of heavy mineral concentration). The heavy mineral constituents of the carbonatite needed to be identified as an aid to interpreting the source of heavy minerals in surficial materials of the northern Superior area (Stone 2001). Heavy minerals separated from the Carb Lake carbonatite include mica (phlogopite and an unknown green mica), two varieties of amphibole including richterite, apatite, barite, fluorite, magnetite, sulphide minerals including pyrite, chalcopyrite, sphalerite, galena and pyrrhotite as well as pyrochlore and synchysite. The heavy mineral separates are stored at the Geoscience Laboratories, Geoservices Centre in Sudbury.

Partial geochemical analyses of sövite, reported by Sage (1987), have high concentrations of CaO and CO_2 consistent with abundant calcite. Combined SiO_2, FeO* and MgO range up to 20% and evidently reflect variable proportions of silicate minerals and magnetite. P_2O_5 probably associated with apatite, ranges up to 8 %. The sövite is enriched in LILE (Ba and Sr), Nb (up to 1500 ppm), Zr (up to 3000 ppm) and LREE.

Sage (1987) recommended the Carb Lake carbonatite to prospectors as a potential source of economic concentrations of niobium and rare earth elements and apatite, the latter of which may occur as residual accumulations in leached cavities on top of the intrusion where it subcrops beneath surficial materials. The occurrence of pyrochlore indicates that the Carb Lake carbonatite can also represent a source of tantalum mineralization.

Gabbro (Diabase) Dikes

Three sets of gabbro dikes are identified on the basis of aeromagnetic anomalies and rare outcrops in the northern Superior area. Set 1 is represented by an *en echelon* series of dikes that locally merge into a single dike and extend north-northeasterly from west of Ponask Lake through Stull and Kistigan lakes (*see* map in back pocket). This dike set is distinguished by a strong positive magnetic anomaly (ODM-GSC 1967b) and is correlated on the basis of orientation with the Molson Swarm, which has an age of 1884 Ma (Krogh et al. 1987). Where exposed west of Ponask Lake and at Stull Lake, the dike has a width in excess of 30 m and is made up of coarse, black weathering brown diabase and has sharp chilled contacts. It is composed of plagioclase+augite+diopside with accessory ilmenite, pyrite and zircon. Pyroxene grains are locally altered to edenite.

Gabbro dike set 2 extends northwesterly through the central northern Superior area (see map in back pocket). Dikes of set 2 have a strong positive aeromagnetic anomaly (ODM-GSC 1967a, b) and are locally curved extending up to 100 km in length. Dikes of set 2 are correlated on the basis of orientation with the MacKenzie Swarm, members of which have a U-Pb age of 1267 Ma (LeCheminant and Heaman 1989). Where exposed south of Swan Lake and at Tamblyn Lake, the dikes of set 2 are represented by brown weathered diabase. The dikes are in excess of 10 m wide, however contacts are not exposed. Samples of dikes of set 2 contain a mineral assemblage of plagioclase+bronzite+diopside with accessory ilmenite and pyrite. Dikes of set 2 appear to be less altered than those of set 1.

Gabbro dikes of set 3 are poorly exposed but appear to be confined to the Northern Superior Superterrane. The dikes are distinguished by moderate to weak negative aeromagnetic anomalies (ODM-GSC 1967a, c) and extend northerly to north-northwesterly. Sample 97DS60 from eastern Yelling Lake and sample 99DS136 from north of Dadson Lake are possibly representative of set 3. The sample

97DS60 contains a mineral assemblage of plagioclase+clinopyroxene+biotite+ilmenite+pyrite. The age of the dikes of set 3 is unknown but is presumably Proterozoic due to the fresh appearance of the rock.

Four samples of dikes of set 2, two samples of set 3 dikes and one sample of an Archean gabbro dike in the Ellard Lake greenstone belt were analyzed geochemically (*see* file *North Superior Geochemistry 2000* on MRD 135). All but the Archean gabbro (a high-Mg basalt) are high-iron tholeiitic basalt and andesite (Figure 18a). The dike rocks are of the medium- to high-K type (Figure 18b) and are strongly metaluminous (ANCK=0.6 to 0.9). The Mg#s are 0.61 for the Archean dike and range from 0.38 to 0.55 for Proterozoic dikes, although most values fall in the range of 0.44 to 0.48. Whereas the Archean gabbro has a flat multi-element profile with slight depletion in incompatible elements, the Proterozoic dikes of sets 2 and 3 are variably enriched in Th and LREE (Figure 18c).

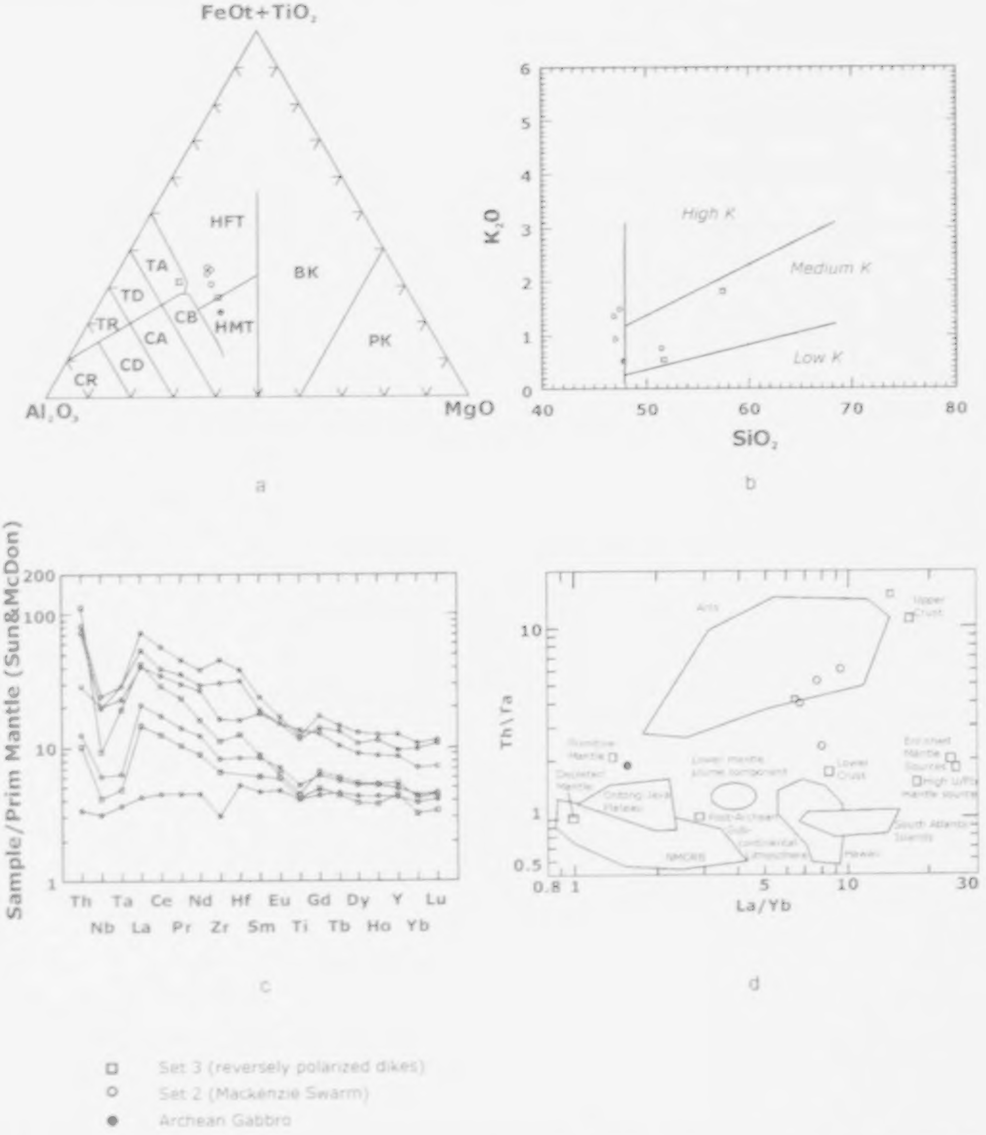


Figure 18. Geochemical plots for mafic dikes. (a) cation plot of Jensen (1976), abbreviations as in Figure 6; (b) SiO$_2$-K$_2$O plot; (c) multi-element plot; and (d) Th/Ta-La/Yb plot of Condie (1997).

Condie (1997) developed a discrimination diagram for the sources of mafic magmas based on Th/Ta and La/Yb relations as shown in Figure 18d. Condie (1997) noted that dikes of the MacKenzie Swarm from elsewhere in the Canadian Shield plot close to the field characteristic of lower mantle plume magmas (*see* Figure 18d) although some dikes have higher Th/Ta and La/Yb due to crustal contamination. In comparison, the Archean gabbro dike from the Ellard Lake greenstone belt plots close to the composition of primitive mantle. The closest modern compositional analogues to the Archean dike are lavas erupted in the Ontong-Java plateau. The Proterozoic dikes have much higher ratios of La/Yb and Th/ Ta than the Archean dike. The evolution of the Proterozoic dikes can perhaps be explained by basalts originating in lower mantle plumes and undergoing extensive contamination by upper crust. This interpretation is supported by the scattering of dikes of both sets 2 and 3 between the fields of lower mantle plume and upper crust (*see* Figure 18d). On the basis of the limited data, there does not seem to be a major compositional difference between dikes of set 2 and those of set 3.

Metamorphism

Bedrock of the northern Superior area consists mainly of plutonic rocks whose mineral assemblages are not sensitive indicators of metamorphic grade. Hence, the definition of metamorphic conditions depends on the sporadic occurrence of units whose mineral assemblages and compositions change systematically with metamorphic grade. The restricted occurrence of informative assemblages, in combination with poor exposure has greatly limited the measurement of metamorphic grade in the area.

Mineral assemblages of prime use for metamorphic studies occur mainly in greenstone belts and include a range of rocks of mafic, ultramafic, pelitic and altered composition. Mineral assemblages representative of the various greenstone belts in the northern Superior area are summarized in Table 7. Generally, the sensitive metamorphic indicator minerals including garnet, aluminosilicates, cordierite and staurolite that are widely used to define metamorphic zones (e.g. Figure 1 of Berman, Easton and Nadeau 2000) are rare in the northern Superior area. This has further restricted the definition of metamorphic zones and application of thermobarometers.

Mafic to ultramafic rocks of the Ponask Lake greenstone belt show a range of mineral compositions representative of amphibolite and greenschist facies (Table 7). Field observations indicate that mafic rocks at the margins of the greenstone belt and in thin greenstone slivers have hornblende-bearing assemblages represented by minerals including hornblende, biotite and feldspars. Greenschist-facies assemblages, represented by minerals including talc, chlorite, albitic plagioclase and local epidote tend to occur at the center of wide parts of the belt.

Mafic volcanic rocks representative of the Stull assemblage in the Stull–Swan lakes greenstone belt have mainly hornblende-bearing mineral assemblages as recognized by field observations and sparse petrographic determinations (Table 7). Mafic volcanic rocks also occur widely in the Ellard Lake and Yelling Lake greenstone belts and appear to contain only hornblende-bearing mineral assemblages. Application of the Al-in-hornblende barometer (discussed later) to coarse recrystallized phases of these mafic sequences yields P-T conditions of 5.0 kbars at 716°C and 4.8 kbars at 715°C at Stull Lake and in the eastern Yelling Lake greenstone belt, respectively (*see* samples 98DS65 and 98DS232 of Table 7).

Hornblende-dominated assemblages in mafic greenstone sequences contrast with assemblages of chlorite, sericite, albitic plagioclase and calcite within mafic and calc-alkalic to alkalic rocks of the Oxford Lake volcanic assemblage at Stull Lake. Notably, sample 98DS64b of pillowed mafic lavas of the Oxford Lake volcanic assemblage at central Stull Lake has a greenschist-facies assemblage of chlorite+albite+epidote. Locally, the volcanic rocks of various mafic assemblages and the Oxford Lake volcanic assemblage are altered and mineralized with sulphide minerals (e.g., samples 96DS300 and 99DS93 of Table 7). The alteration zones contain chlorite with or without garnet and appear to have been metamorphosed to at least upper greenschist facies.

Clastic sedimentary rocks representative of the Cross Lake assemblage in the Stull–Swan lakes greenstone belt have variable mineral assemblages. The matrix of coarse conglomerate at Stull Lake (samples 96DS323b, 96DS324b, 96DS326b) have greenschist-facies assemblages of plagioclase+quartz+chlorite+sericite and may contain biotite and calcite. Boulders of hornblende tonalite within these sedimentary rocks also show the effects of low-grade metamorphism that has converted hornblende to actinolite and chlorite. Eastward within thin greenstone slivers at Swan Lake, two samples of wacke of the Oxford Lake sedimentary assemblage have a mineral assemblage of garnet+biotite+amphibole+plagioclase and provide rare opportunities for application of thermobarometers. The compositions of coexisting minerals were determined by microprobe and used to calculate the P-T conditions with the TWQ software (Berman 1988, 1991). The results for sample 00DS153 are 8.1 kbars at 749°C and for sample 00DS58 are 2.8 kbars at 630°C.

Within broad parts of the Ellard Lake greenstone belt such as between Ellard Lake and Sherman Lake (*see* map in back pocket) hornblende-bearing mafic volcanic rocks tend to occur at margins of the belt whereas intermediate to felsic volcanic rocks and sedimentary rocks occur along the central axis of the belt. In this area, field observations and a petrographically-determined assemblage of ankerite+biotite+plagioclase from sedimentary rocks of the Oxford Lake sedimentary assemblage (sample 98DS45) suggest that a metamorphic grade somewhat lower than amphibolite facies may occur at least sporadically at the center of the Ellard Lake greenstone belt.

The northern Superior area is transected by broad fault zones including the Stull–Wunnummin and South Kenyon and North Kenyon faults. Field observations indicate that faulting occurred by ductile deformation (mylonitization) and brittle deformation (cataclasis). Mylonites are developed in faulted plutonic rocks and are typically crosscut by narrow cataclastic zones and fractures filled with epidote, adularia and prehnite. Faulted mafic volcanic rocks are strongly schistose and appear to contain assemblages of actinolite and/or chlorite. The crosscutting relations suggest that faulting occurred initially by quasi-plastic mylonitization and subsequently by friction-controlled slip, brittle fracture, rigid body rotation and cataclasis. The transition from ductile to brittle deformation mechanisms is largely temperature controlled and occurs at 250 to 350°C corresponding with greenschist conditions (Tullis and Yund 1977; Sibson 1982). This evidence suggests that faulting occurred as metamorphic grade decreased through greenschist facies in the northern Superior area.

No charnockite (orthopyroxene-bearing plutonic rocks), pyroxene-bearing supracrustal rocks or kyanite were observed in outcrop. This may be an indication that “dry” granulite facies metamorphic conditions and high-pressure metamorphism are absent from the study area, however these metamorphic rocks and minerals may have been overlooked due to lack of outcrop. Stone (2001) noted that kyanite, spinel, corundum and orthopyroxene of probable metamorphic origin occur widely in heavy mineral concentrates of surficial materials from the northern Superior area. The occurrence of high-grade metamorphic minerals in surficial materials is an indication the high-grade metamorphic domains undetected by bedrock mapping may occur within or near to the area.

Table 7. Mineral assemblages by greenstone belt, northern Superior area.

Sample No	Area	UTM east	UTM north	Rock Type	Mineral Assemblage
Ponask Lake greenstone belt					
96DS93	Sachigo Lake	563600	5964300	4a	Tlc+Chl+Mgs+Mag+Chr+Py+Pn
97DS42	S Pierce	502980	5996100	5a	Chl+Bt+Grt+Pl+Qtz+Zrn+Py+Ccp+Pn
97DS53	W Pierce	494900	5996800	5p	Amp+Chl+Ilm
97DS45	S Pierce	503650	5996600	7k	Cal+Ank+Tr
Stull-Swan lakes greenstone belt					
Stull assemblage					
96DS300	Gilleran Lake	536000	6021000	5a	Chl+Zo+Bt+Qtz+Kfs+Pl+Grt+Staur+Py+Sph+Ilm
98DS65	Stull L	520500	6025000	5p	Hbl+Pl+Ttn+Py+Ccp. Al in Hbl: P=5.0 kbars, T=715 C
Oxford Lake volcanic assemblage					
98DS64b	Stull L	527700	6031700	5p	Chl+Alb+Ep+Mag+Cal
98DS66	Stull L	524000	6030600	9m	Qtz+Chl+Ser+Alb+Mag+Py+Co-Ni-Arsenide
99DS84	Stull L	529300	6031600	9p	Ser+Chl+Carb+Feld+Mag+Ccp/Py
99DS93b	Rapson Bay	532730	6028700	9t	Qtz+Chl+Sp+Py/Po+Gn+unknown accicular (altered)
99DS97	Stull L	529510	6028660	9n	Ser+Act+Carb+Qtz+Feld+Mag+Py/Ccp
99DS303	Rapson Bay	531250	6030960	9m	Pl+Chl+Cal+Mag/Ilm+Py+green/brown amph?
Oxford Lake sedimentary and Cross Lake assemblages					
96DS323b	Stull Lake	524000	6031000	9c	Pl+Qtz+Chl+Ser+Cal+Ep+Mag
96DS324b	Stull Lake	524000	6031000	9c	Pl+Qtz+Chl+Ser+Mag+Ep
96DS326b	Stull Lake	524000	6031000	9c	Pl+Qtz+Chl+Ser+Bt+Cal+Mag+unknown
99DS224	W. of Meston	574619	6015841	9b	Qtz+feld+ser+Cal (clastic sediment)
99DS227	SW of Meston	576426	6012906	9b	Grt+Amp+Pl+Qtz+Py+Mag+Chl
00DS58	Swan	617950	6013450	9b	Grt+Bt+Amp+Pl+Qtz+Py+Ep/Zo TWQ P=2.8 kbars, 630 C
00DS153	N Rawley Lk	621205	6007000	9b	Grt+Amp+Pl+Qtz+Bt+Carb+Mag/Ilm TWQ P=8.1 kbars, T=749 C
Ellard Lake greenstone belt					
98DS31	Ellard L	568389	6045772	5a	Amp+Chl+Pl+Mag+Rt+Py
98DS125	Lacey L	582600	6039600	5p	Hbl+Pl+Ttn+Chl+Ap+Kfs+Py
98DS56	Foster L	603100	6041800	5a	Bt+Hbl+Pl+Zo+Ttn
98DS45	Gummer L	598700	6039900	9a	Qtz+Ank+Bt+Pl+Kfs+Py+Mag+Ilm
Yelling Lake greenstone belt					
97DS70	E of Yelling	590800	6074300	5g	Amp+Pl+Bt+Ttn+Qtz+Chl+Ilm+Carb(vein)
98DS331	E of Shallow	626400	6069900	5a	Hbl+Ep+Qtz+Mag+Pl
98DS232	N. Dadson	615177	6068800	5a	Amp+Pl+Ep+Ttn+Py+Cp. Al in Hbl: P=4.8 kbars, T=715 C

Mineral abbreviations are from Kretz (1983); Amp=amphibole; Carb=carbonate

Al in Hbl- Aluminum in hornblende barometer

TWQ-thermobarometer of Berman (1988, 1991)

THE AGE OF METAMORPHISM

In large greenstone belts such as at Stull Lake, basalts of the various mafic assemblages are typically at amphibolite grade of metamorphism whereas volcanic and sedimentary rocks of the Oxford Lake assemblage are at greenschist facies. This may be an indication that the area was affected by an early and possible pre 2800 Ma metamorphic event comparable to the Wachusk metamorphic domain in the North Caribou terrane (Easton 2000). This early metamorphism would have caused amphibolite-facies metamorphism in older mafic assemblages but would not be recorded in the younger rocks of the Oxford Lake assemblage. Supracrustal rocks of the Oxford Lake assemblage show mainly a lower grade metamorphism comparable to P_1 metamorphism (Easton 2000), which affected the western Superior Province from 2710 to 2695 Ma and may have been a few million years earlier in the north (the present area).

On the other hand, the Oxford Lake assemblage tends to occur in the central parts of the Stull–Swan lakes greenstone belt (*see* map in back pocket) and the variation in metamorphic grade between the mafic assemblages and the Oxford Lake assemblage may reflect metamorphic zonation of greenstone belts. Ayres (1978) noted that large greenstone belts of the western Superior Province tend to be metamorphically zoned with greenschist-facies cores and amphibolite-facies rims. This metamorphic pattern was interpreted by Ayres (1978) to indicate that greenstone belts were affected by regional greenschist-facies metamorphism and the rims were metamorphosed to amphibolite facies by the thermal contact effects of surrounding batholiths.

Some field evidence and geochronology supports the latter of the above interpretations. For example, the Ponask Lake greenstone belt is composed of pre 2800 Ma mafic volcanic strata and has greenschist-facies metamorphic rocks in central parts of the belt (e.g., sample 96DS93 of Table 7). This metamorphic pattern is not consistent with pre 2800 Ma regional metamorphism to amphibolite facies. Further, rocks of the Oxford Lake assemblage show amphibolite-facies mineral assemblages where they occur in thin and faulted greenstone slivers in the Swan Lake area (samples 00DS58 and 00DS153, Table 7). The variation in metamorphic grade for rocks of the Oxford Lake assemblage from low-grade in thick greenstone sequences to high-grade in thin sequences is best explained by post 2722 Ma contact metamorphism possibly augmented by block faulting at Swan Lake.

Scant geochronological data on metamorphic age is derived from U-Pb zircon analyses of plutonic rocks (*see* Table 3). Skulski et al. (2000) noted thin, unzoned and discontinuous rims on zircon grains from tonalitic and gneissic rocks of the Island Lake terrane and Northern Superior Superterrane. The rims were interpreted as metamorphic overgrowths on cores of magmatic origin. SHRIMP analyses gave ages of 2758, 2741 and 2717 Ma for the rims and pre 2800 Ma ages for the cores of the zoned grains (*see* Table 3). These authors also noted that 2714 Ma zircon grains of the sanukitoid suite lack metamorphic rims and suggested that metamorphism of the Northern Superior Superterrane must have occurred in the interval 2741 to 2714 Ma. Ketchum (2002) obtained an average age of 2715 Ma on 3 fragments that he interpreted to be spalled-off metamorphic overgrowths on zircon from a 2725 Ma plagioclase porphyry at Swan Lake (No. 26 of Table 3).

Collectively, the field relationships and geochronology can be explained by one or more Neoarchean metamorphic events although earlier metamorphism cannot be ruled out. Metamorphism appears to have occurred earlier in the northern Superior area than in southern parts of the western Superior Province (pre 2714 Ma vs. post 2710 Ma; Easton 2000). Variations in metamorphic grade within the area can be accounted for by a combination of contact metamorphic effects and late block faulting, the latter of which is further resolved by Al-in-hornblende barometry as discussed below.

Aluminum-in-Hornblende Barometry

Aluminum- or Al-in-hornblende barometry takes account of the linear relation between pressure and the Al-content of hornblende and is widely used to estimate the crystallization pressure (depth of crystallization) for intermediate to felsic igneous rocks containing an appropriate buffer assemblage that includes hornblende. The Al-in-hornblende barometer was originally defined on an empirical basis by Hammarstrom and Zen (1986) and subsequently modified using larger data sets by several workers including Schmidt (1992). Blundy and Holland (1990) showed that temperature also affects the Al-content of hornblende and these authors produced a hornblende thermometer that they subsequently modified (Holland and Blundy 1994) on the basis of a larger experimental data set.

Application of the early Al-in-hornblende barometers carries the implied assumption that temperature is invariant in vicinity of the solidus temperature, which is in the range of 685°C to 650°C for H_2O-saturated magmas of tonalite to granodiorite composition (Hollister et al. 1987). Anderson and Smith (1995) showed that the solidus temperatures for plutonic magmas can exceed 700°C and went on to develop a temperature-corrected barometer. These authors also showed that amphiboles crystallized under low oxygen fugacity give artificially high pressures and proposed chemical tests based on the Fe/Mg ratios of amphiboles to screen anomalous pressure determinations. The interested reader is referred to Stone (2000) for a review of the appropriate conditions for application of Al-in-hornblende barometry.

Equations 2, 3 and 4 of Stone (2000) are applied to analyses of amphibole+plagioclase pairs in the present study and results (P_{Schmidt}; $T_{\text{Holland and Blundy}}$; $P_{\text{Anderson and Smith}}$) are shown with the amphibole analyses in the data file *Amphibole* on MRD 135. These equations represent the barometer of Schmidt (1992), the thermometer of Holland and Blundy (1994) and the temperature-corrected barometer of Anderson and Smith (1995), respectively. In approximately a third of the samples, a temperature-corrected pressure could not be obtained ($P_{\text{Anderson and Smith}}$=unsuitable) because amphibole analyses failed the chemical tests of Anderson and Smith (1995).

RESULTS

The majority of PT determinations by the Al-in-hornblende method were obtained from hornblende-bearing plutonic rocks of the hornblende tonalite and sanukitoid suites. A few PT determinations were made from other plutonic suites including the biotite tonalite, biotite granite and gneissic suites at localities where amphibole occurs as an auxiliary mineral phase. The Al-in-hornblende barometer was also applied to a few coarse recrystallized samples of mafic volcanic rocks, gabbro and a gabbroic phase of a MacKenzie dike.

Temperature and pressure data are grouped according to rock type, crustal block and age of the rock in Tables 8 and 9. Crustal blocks are defined partly on the basis of Al-in-hornblende pressures and are large domains of fairly uniform-P and are typically bounded by major faults or greenstone belts.

TEMPERATURE

Crystallization temperatures were obtained from 92 samples distributed widely in the northern Superior area. The temperatures range from 630°C to 749°C with an average of 699°C. The extreme temperatures are obtained from the Oxford Lake sedimentary assemblage using the garnet-biotite exchange thermometer in the program TWQ of Berman (1988; 1991). T-estimates are up to 100°C higher than wet solidus temperatures for most felsic plutonic rocks. This indicates that crystallizing magmas were not water-saturated and justifies use of the temperature-corrected barometer.

Excluding samples 00DS58 and 00DS153 of the Oxford Lake sedimentary assemblage, temperatures are somewhat higher in the biotite tonalite, tonalite gneiss and biotite granite suites than in the hornblende tonalite and sanukitoid suites. This may simply be an indication that the former magmas were somewhat drier than the latter. The preponderance of the relatively old biotite-bearing plutonic suites in the Island Lake terrane causes a trend of higher temperatures in pre 2800 Ma rocks and in crustal block 1 (Table 8). Below-average temperatures were obtained from mafic rocks including mafic volcanic rocks, gabbro and the gabbroic phase of a MacKenzie dike. The samples of the mafic rocks were obtained from coarse, recrystallized phases possibly representing late deuteric zones that may account for the low temperatures.

Table 8. Temperature data from the northern Superior area.

Rock Type[1]	**Age (Ma)**[2]	**Crustal Block**[3]	**T °C[4] ±40°C**	**No. of Samples (total=92)**
Temperature variation by rock type				
5	2857?, 2732 to 2718	1,2,3, 4	680	9
9s (samples 00DS153, 00DS58)	2722?	2	749, 630	1
10	2712	2	638	4
11, 12	2859 to 2822, 2738	1, 2, 4	728	10
15	2722, 2723	1, 2, 4	719	5
16	2716	1, 2, 3, 4	708	48
14	2712	3, 4	670	14
30	1267	1	650	1
Temperature variation by crustal block				
12, 15, 16	Variable	1 (excludes rock type 30)	724	24
5, 7, 10, 12, 15, 16	Variable	2	689	18
14, 16	Variable	3	691	26
11, 14, 16	Variable	4	689	24
Temperature variation by age of rock				
5, 11, 12	2857 to 2822	1, 2	730	10
12	2738 to 2732	2	653	6
5, 7, 10, 15	2723 to 2722	1, 2	695	10
5	2718?	4	703	2
16	2716	1, 2, 3, 4	708	48
10, 14	2712	2, 3, 4	673	15
30	1267	1	650	1
Overall average of Archean rocks	**Variable**	**1, 2, 3, 4**	**699**	**91**

[1] Rock types: 5-mafic volcanic; 9s-sediments; 10-gabbro, diorite associated with 14; 11- tonalite gneiss; 12- biotite tonalite; 14-monzodiorite (sanukitoid); 15-biotite granite; 16-hornblende tonalite; 30-McKenzie gabbro dike.

[2] Age is an average of measured ages for rocks of this plutonic suite or the supracrustal assemblage from which the sample is thought to belong.

[3] Crustal Blocks: 1-south of the Stull-Wunnummin fault; 2-north of the Stull-Wunnummin fault and south of the Ellard Lake greenstone belt; 3-north of the Ellard Lake greenstone belt and south of the South Kenyon Fault; 4-north of the South Kenyon Fault.

[4] T°C Temperature in degrees C by the method of Holland and Blundy (1994); samples 00DS58 and 00DS153 by garnet+hornblende+biotite+plagioclase equilibria.

PRESSURE

Temperature-corrected pressures were obtained from 67 samples in the northern Superior area and range from 1.0 kbars to 9.6 kbars with an average of 4.4 kbars (see $P_{\text{Anderson and Smith}}$ with the data file *Amphibole* on MRD 135). Pressure determinations were made on all plutonic suites except for the two-mica granite suite and are distributed widely in the northern Superior area (Figure 19a). The majority of pressure determinations were obtained from hornblende-bearing plutonic suites including the hornblende tonalite suite and sanukitoid suites.

Pressure data is grouped according to rock-type, crustal block and age of rock in Table 9 where several important trends can be observed. Firstly, average pressures are high in supracrustal rocks and gabbro and show a decreasing trend through biotite tonalite, biotite granite, hornblende tonalite and the sanukitoid suites of plutonic rocks and appear to be lowest in McKenzie gabbro dikes on the basis of a single determination in the latter rock. This trend can be explained by a pattern of decreasing pressure with age of the rock (compare *Pressure variation by rock type* and *Pressure variation by age of rock* in Table 9). Although supracrustal rocks in mafic assemblages can be on the order of 2.85 Ga old, the high pressures in these and younger supracrustal rocks probably represent peak metamorphic conditions that occurred at 2714 to 2741 Ma at least in the Northern Superior Superterrane. Peak metamorphism was followed by emplacement of a succession of plutonic suites with average ages ranging from 2738 Ma to 2712 Ma. Amphiboles crystallized at progressively lower pressures in the younger plutonic suites. This data is interpreted to indicate rapid uplift during the late Archean and uplift appears to have continued at a slower rate into the Proterozoic as indicated by the lowest pressure determination, which is associated with a 1267 Ma McKenzie gabbro dike.

The second significant trend in the pressure data is spatial variation. Pressure determinations for the hornblende tonalite suite are sufficiently abundant and widespread for display by a contoured isogonal map. The map (Figure 19b) shows spatial variation in crystallization pressure at 2716 Ma (the average age of the hornblende tonalite suite). Contours are drawn manually on Figure 19b and outline four major crustal domains or blocks characterized by more-or-less uniform pressure. The blocks, for which pressure data is summarized in Table 9 and whose boundaries are shown in Figure 19c, tend to be demarcated by major faults or greenstone belts.

Table 9. Pressure data from the northern Superior area.

Rock Type[1]	**Age (Ma)**[2]	**Crustal Block**[3]	**P (kbar)±1.0 kbar**[4]	**No. of Samples (total=68)**
Pressure variation by rock type				
5	2857? 2718?	1, 4	4.9	2
9s (samples 00DS153, 00DS58)	2722?	2	8.0	1
10	2712	2	6.6	1
11, 12	2863 to 2822, 2738, 2701	1, 2, 4	5.0	5
15	2723	1, 2	4.5	4
16	2716	1, 2, 3, 4	4.4	42
14	2712	3, 4	2.9	8
30	1267	1	1.0	1
Pressure variation by crustal block				
5, 12, 15, 16	Variable	1 (excludes rock type 30)	4.7	22
7, 10, 12, 15, 16	Variable	2	6.9	9
14, 16	Variable	3	3.8	16
11, 14, 16	Variable	4	3.5	20
Pressure variation by age of rock				
5, 11, 12	2857 to 2822	1, 2	4.6	9
12	2738	2	8.7	1
7, 15	2723 to 2722	1, 2	5	5
5	2718?	4	4.8	1
16	2716	1, 2, 3, 4	4.4	42
10, 14	2712	2, 3, 4	3.3	9
30	1267	1	1.0	1
Overall average of Archean rocks	**Variable**	**1, 2, 3, 4**	**4.4**	**67**

[1] Rock types: 5-mafic volcanic; 9s-sediments; 10-gabbro, diorite associated with 14; 11- tonalite gneiss; 12- biotite tonalite; 14-monzodiorite (sanukitoid); 15-biotite granite; 16-hornblende tonalite; 30-McKenzie gabbro dike.

[2] Age is an average of measured ages for rocks of the plutonic suite or the supracrustal assemblage to which the samples are thought to belong.

[3] Crustal Blocks: 1-south of the Stull-Wunnummin fault; 2-north of the Stull-Wunnummin fault and south of the Ellard Lake greenstone belt; 3-north of the Ellard Lake greenstone belt and south of the South Kenyon Fault; 4-north of the South Kenyon Fault.

[4] P (kbar) pressure in kilobars by the Al-in-hornblende method, samples 00DS58 and 00DS153 by garnet+hornblende+biotite+plagioclase equilibria.

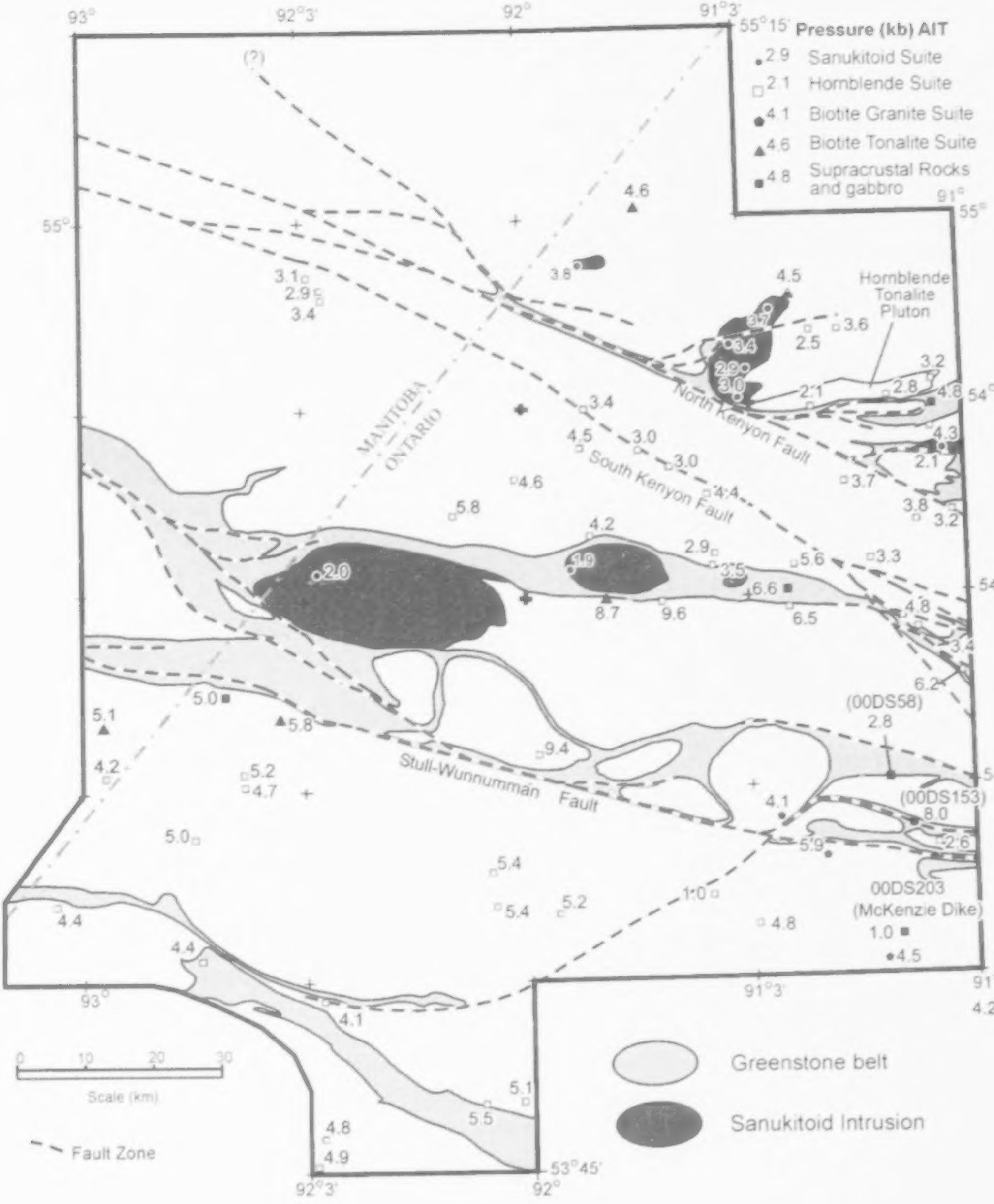

Pressure (kb) AIT
2.9 Sanukitoid Suite
2.1 Hornblende Suite
4.1 Biotite Granite Suite
4.6 Biotite Tonalite Suite
4.8 Supracrustal Rocks and gabbro
Greenstone belt
Sanukitoid Intrusion
Fault Zone
Scale (km)
0
10
20
30
93°
92°3'
92°
91°3'
91°
55°15'
55°
54°45'
54°3'
54°15'
54°
53°45'
MANITOBA
ONTARIO
North Kenyon Fault
South Kenyon Fault
Stull-Wunnumman Fault
Hornblende Tonalite Pluton
(00DS58)
(00DS153)
00DS203 (McKenzie Dike)
(?)

Figure 19a. Aluminum-in-hornblende pressure determinations (kbars) for plutonic suites.

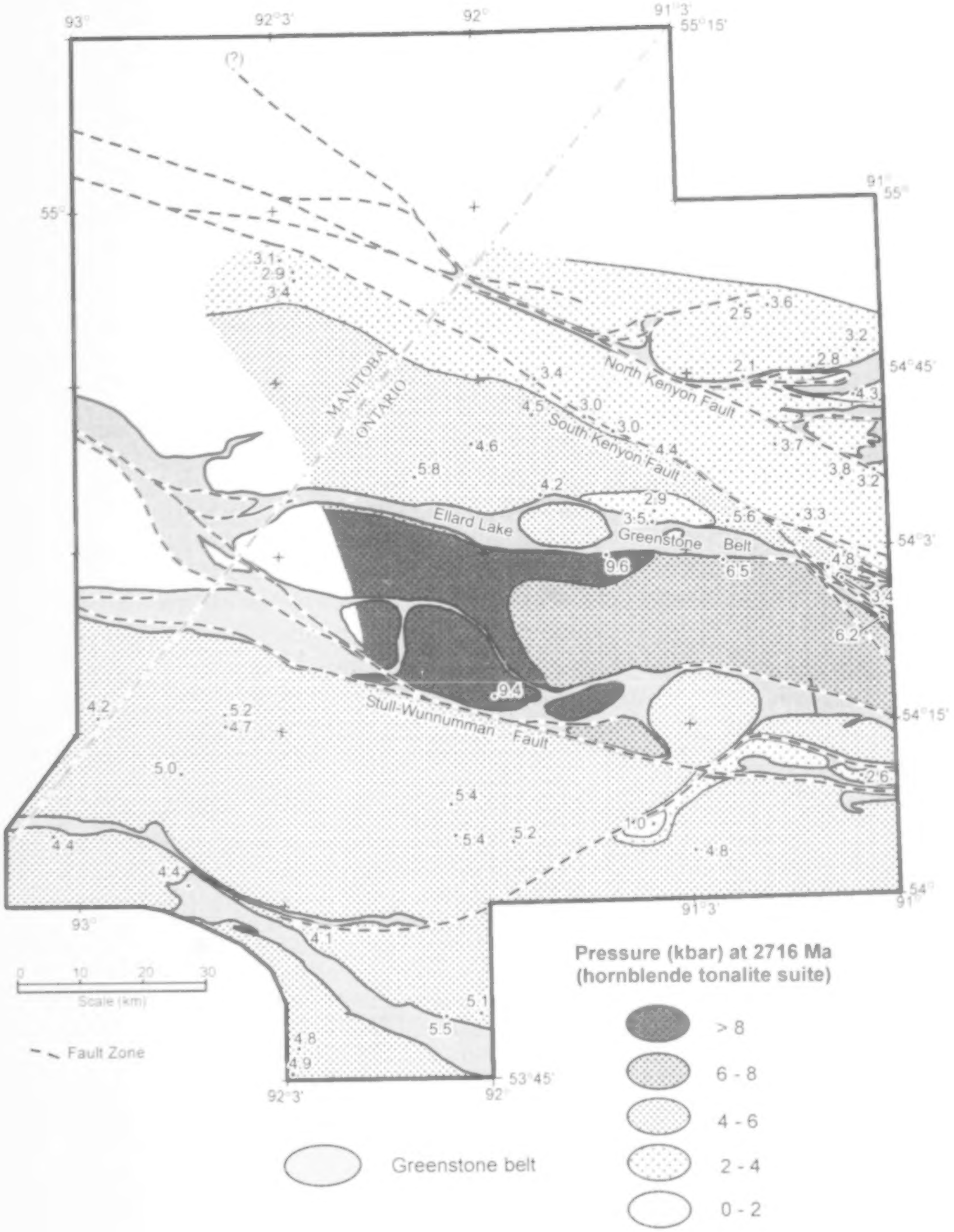

93°
92°3'
92°
91°3'
55°15'
91°
55°
55°
54°45'
54°3'
54°15'
54°
91°
91°3'
93°
92°3'
92°
53°45'
(?)
MANITOBA
ONTARIO
North Kenyon Fault
South Kenyon Fault
Ellard Lake
Greenstone Belt
Stull-Wunnumman Fault
3.1
2.9
3.4
2.5
3.6
3.2
2.1
2.8
4.3
3.4
4.5
3.0
3.0
4.4
3.7
3.8
3.2
4.6
5.8
4.2
2.9
3.5
5.6
3.3
9.6
6.5
4.8
3.4
6.2
9.4
4.2
5.2
4.7
5.0
5.4
5.4
5.2
1.0
4.8
2.6
4.4
4.4
4.1
5.1
5.5
4.8
4.9
0
10
20
30
Scale (km)
Fault Zone
Pressure (kbar) at 2716 Ma
(hornblende tonalite suite)
> 8
6 - 8
4 - 6
2 - 4
0 - 2
Greenstone belt

Figure 19b. Contoured isogonal plot of aluminum-in-hornblende pressure determinations for the hornblende tonalite suite

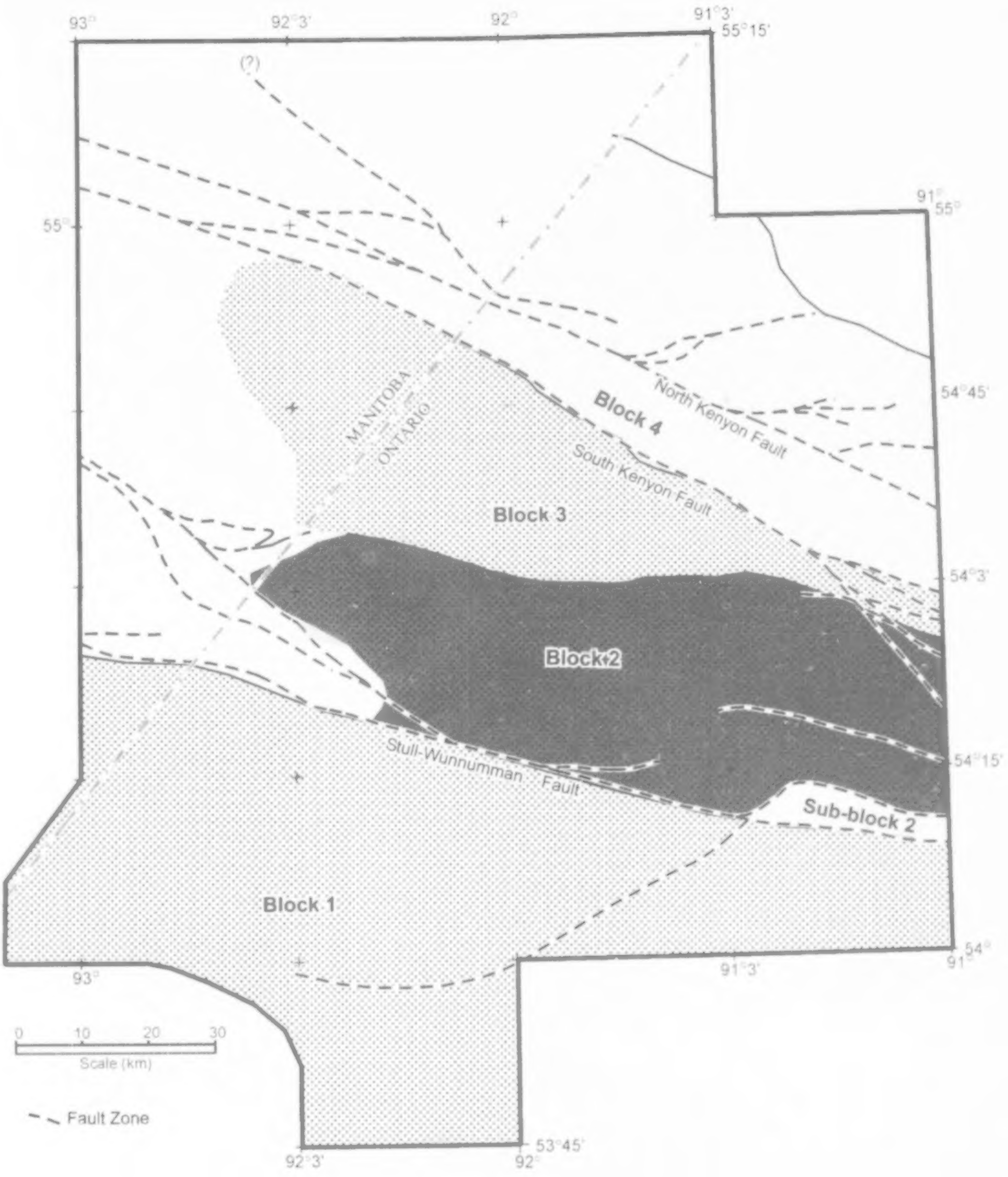

93°
92°3'
92°
91°3'
55°15'
(?)
91°
55°
55°
54°45'
MANITOBA
ONTARIO
Block 4
North Kenyon Fault
South Kenyon Fault
Block 3
54°3'
Block 2
Stull-Wunnumman Fault
54°15'
Sub-block 2
Block 1
54°
91°
93°
91°3'
0
10
20
30
Scale (km)
Fault Zone
53°45'
92°3'
92°

Figure 19c. Crustal blocks of the northern Superior area from aluminum-in-hornblende pressure determinations.

Block 1 occurs south of the Stull–Wunnummin fault corresponding more-or-less with the Island Lake terrane and has hornblende crystallization pressures in the range of 4 to 6 kbars. Block 2 includes parts of the Oxford Lake–Stull Lake terrane situated north of the Stull–Wunnummin fault and south of the Ellard Lake greenstone belt and has high pressures in the range of 6 to 9 kbars. Sub-block 2 is a small, fault-bounded area that possibly represents a domain of low pressure on the basis of one pressure determination. Block 3 represents parts of the Oxford Lake–Stull Lake terrane extending north of the Ellard Lake greenstone belt to the South Kenyon fault and has pressure mainly in the range of 4 to 6 kbars. Overall low pressures less than 4 kbars occur in Block 4, which extends north of the South Kenyon fault and corresponds closely with the Northern Superior Superterrane.

The pressure variations from block to block can be explained by post 2716 Ma dip-slip displacement on bounding faults. Although normal displacements are possible, reverse faulting is favoured because it is consistent with the overall compressional tectonic regime related to accretion of the various crustal blocks. In this scenario, Block 1 has underthrust Block 2 and Block 4 has underthrust Block 3, which has in turn underthrust Block 2 (*see* Figure 4). Evidently, the Ellard Lake greenstone belt must overlie and mask a crustal discontinuity not detected by bedrock mapping. Details regarding the timing and relative vertical displacements of the blocks can be gained by comparing pressure data from the sanukitoid suite with that of the hornblende tonalite suite and has been discussed previously (*see* "Terrane Interaction").

Structure

REGIONAL STRUCTURAL TRENDS

Regional structural trends defined by lithologic contacts, foliations, gneissosity and faults are strongly aligned east-southeasterly to easterly in the northern Superior area (*see* map in back pocket). Variations in structural trends occur locally. For example, foliations and lithologic contacts wrap completely around the oval granite pluton north of Blackbear Lake. Also, several faults extend northeasterly to easterly through the area. These include the fault extending northeast from Ponask Lake to Blackbear Lake and a splay that extends east from the North Kenyon fault near Yelling Lake. Lithologic contacts and foliations tend to be deflected into northeasterly trends within and adjacent to these faults.

Foliations defined by alignment and long axes of minerals have mainly steep to vertical dip. Local areas of shallow to intermediate dip are identified such as at Yelling Lake in the Northern Superior Superterrane. Also, strata dips consistently southward within a strip extending a few kilometres south of the Stull–Wunnummin fault. Otherwise, the area is characterized by a fairly well-developed mineral fabric with overall east-southeasterly trend and sub-vertical dip.

Although some late plutons of the biotite granite and sanukitoid suites have distinct oval shape, most lithologic units are highly elongate parallel to regional structural trends. For example, although greenstone belts locally attain a width of up to 15 km, they are typically of only a few kilometres width and can be extensive for tens of kilometres over which distance they may curve, bifurcate and anastomose but otherwise, maintain a narrow belt-like form. In comparison with major greenstone belts in the southern part of the western Superior Province such as at Birch–Uchi lakes, Lake of the Woods and Sioux Lookout (Ontario Geological Survey 1991a,b), greenstone belts of the northern Superior area tend to be narrow and straight. Likewise, many felsic plutons are elongate east-southeasterly in the northern Superior area.

The combination of strongly preferred orientation of structural trends and lithologic units together with well-developed mineral fabric and major terrane boundary faults implies that the level of strain may be unusually high in the northern Superior area. Three accreted crustal terranes are transected over a northerly-trending distance of 130 km in the northern Superior area. Although terranes are not necessarily narrower in the north than in the south part of the western Superior Province (*see* Figure 3), the terrane boundaries are marked by major faults in the north. Al-in-hornblende barometry demonstrates that the boundary faults have accommodated significant dip-slip displacement related to the flattening component of strain originating from oblique northerly transpressive deformation. The flattening resulted from accretionary assembly of the terranes in the northern Superior Province and provides a possible mechanism for explaining the strong structural trends.

DEFORMATION EVENTS

Jiang and Corkery (1998) provided the only known systematic structural analysis of deformation events within and adjacent to the northern Superior area. These authors identified four deformation events in the Edmund Lake–Little Stull Lake area of Manitoba.

The earliest deformation event (D_1) is characterized by a shallow-dipping foliation and shallow-plunging stretching lineation. D_1 is developed only in the older mafic assemblages of the western Stull–Swan lakes greenstone belt, particularly near the margins of the belt, and is interpreted to be tectonic in origin and possibly developed within low-angle shear zones. Skulski et al. (2000) suggested that D_1 fabric developed due to thrusting of the older mafic oceanic assemblages, such as the Edmund Lake and Hayes River assemblages onto the margin of the Island Lake terrane prior to eruption of the Oxford Lake volcanic assemblage.

Gently plunging, regional isoclinal folds are well developed in the Little Stull Lake area and have been interpreted as originating from the second deformation event (D_2). All supracrustal sequences including the older mafic assemblages and younger strata such as the Cross Lake assemblage are affected by D_2. Jiang and Corkery (1998) noted that the second deformation event is largely responsible for the map pattern in greenstone belts and appears to have originated due to regional flattening deformation. D_2 may also be instrumental in creating the regional structural trends and causing reverse displacement on major faults as discussed above.

Regional-scale shear zones such as the Stull–Wunnummin fault represent the third deformation event (D_3). A variety of kinematic indicators and sub-horizontal mineral lineations show that the shear zones have accommodated primarily dextral transcurrent shearing. The strike-slip shearing is evidently superimposed on earlier dip-slip displacement related to D_2. Jiang and Corkery (1998) suggested that D_3 is a continuation of D_2 in which earlier flattening strain responsible for folding gave way to non-coaxial dextral transpression.

The fourth deformation event (D_4) is recognized by outcrop-scale S-folds in the Little Stull Lake area, some of which refold previously developed mylonite of the Stull–Wunnummin fault. Jiang and Corkery (1998) noted conjugate pairs of S-folds, which they attributed to north-south shortening possibly related to emplacement of granite plutons at margins of the greenstone belt.

STRUCTURAL YOUNGING DIRECTIONS AND FOLDS

The local younging direction or "way up" of supracrustal strata is measured on the basis of grain size gradations in sedimentary beds and from the shapes of pillows in volcanic sequences, in several greenstone belts of the northern Superior area. As discussed by Borradaile (1976), the local younging direction can change considerably within folded strata. The younging direction does not necessarily represent the regional structural facing direction (the component of the younging vector projected onto the axial plane cleavage) but nonetheless provides useful information on the structure of greenstone belts and the stratigraphic relations between various units within the belts.

Pillows within mafic sequences in north-central parts of the Ponask Lake greenstone belt at Sachigo Lake young southward at three localities (see map in back pocket). These are consistent with a southward younging direction inferred from stratigraphy at Ponask Lake where tonalite cobble conglomerate occurs south of a tonalitic basement complex.

Likewise, pillows within mafic sequences of the Pierce assemblage 10 km east of Pierce Lake young southward. These data suggest that strata at least along the north margin of the Ponask Lake greenstone belt youngs south.

The structure of the Stull–Swan lakes greenstone belt at Stull Lake and Little Stull Lake is complex and only partly understood. Pillows within the mafic sequences represented by the Stull assemblage at the south side of the belt and the Edmund Lake and Rorke Lake assemblages at the north side of the belt young northward. Although undated, the mafic sequences appear to represent distinct early structural panels composed of overall northerly younging strata that occur in fault contact with younger rocks at the central axis of the belt.

The central part of the Stull–Swan lakes greenstone belt at Stull Lake is represented by alternating units of the Oxford Lake volcanic and sedimentary assemblages and the Cross Lake assemblage. The clearest structural relations occur at south-central Stull Lake where the Oxford Lake volcanic assemblage is dated at 2723 Ma and youngs south where it is unconformably overlain by conglomerate of the Cross Lake assemblage. The conglomerate also youngs south giving way to coarse sandstone, which is dated at <2709 Ma. These structural and geochronologic data suggest an overall southward-younging stratigraphic package at south-central Stull Lake. At northern Stull Lake and Rapson Bay, the Oxford Lake and Cross Lake assemblages are highly faulted and show northerly and southerly younging directions. The structure of this area may be represented by a series of fault-bounded panels or isoclinally folded strata.

At Little Stull Lake, Corkery, Skulski and Whalen (1997) identified four, largely fault-bounded structural panels. Panel 1 represents mafic sequences south of the "Wolf Bay" shear zone (a splay of the Stull–Wunnummin fault corresponding to D_3 of Jiang and Corkery 1998). Panel 1 is correlative with the north-younging northern panel of Edmund Lake and Rorke Lake assemblages at Rapson Bay of Stull Lake. Panel 2 occurs as a wedge of fluvial-alluvial sandstone and conglomerate at eastern Little Stull Lake. The sedimentary rocks occupy a west-facing anticline (D_2 fold of Jiang and Corkery 1998) that is cored by arkosic sandstone overlain by conglomerate. Panel 3 is a complexly folded and fault-bounded package of west-facing intermediate to felsic volcanic rocks of the Oxford Lake volcanic assemblage at northern Little Stull Lake. Panel 4 represents mafic sequences of the Rorke Lake assemblage within a narrow unit at southwestern Kistigan Lake. Although no structural younging data was obtained from this panel, it probably represents mafic sequences thrust during D_1 deformation onto the margin of the Island Lake terrane.

In the area between Ellard Lake and the Sachigo River, local younging directions are derived from pillow shapes in mafic sequences of the Ellard assemblage. Younging directions are mainly southward along the north side of the Ellard Lake greenstone belt and northward along the south side of the belt. These structural data imply that the Ellard Lake greenstone belt has a broadly synclinal form in this area. A narrow, bifurcated unit of the Oxford Lake volcanic assemblage occupies the core of the syncline.

The Yelling Lake greenstone belt is characterized by a series of narrow and highly strained greenstone slivers. Reliable younging directions are rarely observed in this belt.

In summary, structural analysis of greenstone belts in the northern Superior area is plagued by poor exposure and high levels of strain that obliterate primary features (e.g., pillows and graded bedding) used to obtain younging directions. Nonetheless, the larger belts such as at Stull Lake seem to be composed of fault-bounded structural panels. Mafic panels typically show sections of up to a few kilometres in thickness characterized by consistent younging direction. These may represent homoclinal sections or tectonically thickened sections. Locally, such as east of Ellard Lake, the mafic sequences are folded, however fewer folds are recognized in the older mafic sequences than in the younger sequences.

Younging directions are more variable in intermediate to felsic volcanic and sedimentary sequences than in mafic sequences at Stull Lake and Little Stull Lake. The variable younging directions suggest that regional-scale folding correlative with the D_2 event of Jiang and Corkery (1998) has been preferentially concentrated in the Oxford Lake and Cross Lake assemblages. D_2 folding is thought to have developed during the flattening stages of terrane interaction. Possibly, the mafic sequences were mechanically rigid and resisted folding, with the result that folds preferentially developed in the less competent felsic volcanic and sedimentary sequences.

FAULTS

The northern Superior area is cut by three major faults and numerous, subsidiary splay faults. The major structures include the Stull–Wunnummin, South Kenyon and North Kenyon faults (see map in back pocket). These structures have been traced using airborne geophysical data for hundreds of kilometres east-southeasterly (Osmani and Stott 1988). The major faults are interpreted to represent terrane boundaries with the Stull–Wunnummin fault separating the Island Lake and Oxford Lake–Stull Lake terranes. The South Kenyon fault demarcates the Oxford Lake–Stull Lake terrane from the Northern Superior Superterrane and the North Kenyon fault appears to lie entirely within the Northern Superior Superterrane.

Although poorly exposed, the major faults contain a variety of fault rocks that record early stages of ductile deformation followed by brittle deformation. Where the faults cut plutonic rocks, mylonites are developed and typically show rounded feldspar augen within a strongly foliated, fine-grained matrix of quartz and feldspar. The mylonites are locally cut by narrow brittle faults containing finely comminuted cataclastic material. Fractures filled with epidote and chlorite typically mark the latest stages of brittle deformation. Based on the superposition of cataclastic features on mylonites, it can be concluded that faulting protracted over a period of time during which fault rocks cooled and embrittled. Where the faults cut supracrustal rocks such as at Stull Lake, schists are developed and primary features including pillows and cobbles are highly stretched. Chlorite and sericite are common minerals within faulted supracrustal rocks indicating that deformation occurred under greenschist conditions. The schists are locally cut by veins filled with quartz and carbonate.

Jiang and Corkery (1998) noted mainly sub-horizontal mineral lineations and kinematic indicators representing a dextral sense of shear in the Stull–Wunnummin fault. They suggested that the Stull–Wunnummin fault had accommodated mainly dextral transcurrent displacement. The South and North Kenyon faults also show a range of intermediate- to shallow-plunging mineral lineations that are consistent with overall transcurrent displacement. Mylonites within the South and North Kenyon faults show a preponderance of Z-shape over S-shape minor folds and markers are offset dextrally by late brittle faults; these kinematic indicators are consistent with a dextral sense of shear. The dextral transcurrent displacement on major faults (D_3) represents the later transpressive stage of terrane amalgamation following the earlier D_2 flattening stage.

Although structural features show that the latest stages of deformation on the major faults involved dextral transcurrent motion, results of Al-in-hornblende barometry summarized in Figure 4 record significant vertical motion of crustal blocks. The relative vertical motion of crustal blocks has been explained by thrust displacement on the boundary faults. Thrusting, associated with the D_2 flattening stage of terrane abutment, occurred at 2716 to 2714 Ma on the South Kenyon fault and after 2710 Ma on the Stull–Wunnummin fault (*see* "Terrane Interaction"). Although the D_3 transcurrent motion probably marked a continuation of the thrusting, perhaps through intermediate stages of oblique-slip motion, the transcurrent motion protracted longer and became the final stage of faulting as shown by kinematic indicators. Parks et al. (2003) obtained an age of 2700 Ma from a weakly boudinaged dike cutting the shear-zone fabric of the Savage Island Shear Zone at Island Lake. These authors concluded that the final stages of shearing on the Savage Island Shear Zone are represented by the 2700 Ma date. Broadly, 2700 Ma may mark the time of transcurrent faulting in the northern Superior area.

Economic Geology

Owing largely to remoteness and poor exposure, the northern Superior area has had little previous mineral exploration in comparison with southern parts of the Superior Province. Early prospecting for gold led to discovery of the Sachigo River Mine, which represents the only mineral production from the area. Parts of the northern Superior area were explored sporadically for base metals and gold in the 1970s and 1980s (Assessment Files, Resident Geologist's Office, Red Lake). Recently, exploration has focused on gold in the Stull–Swan lakes greenstone belt at Little Stull Lake and Twin Lakes, Manitoba.

Despite rather little exploration work, the northern Superior area has a wide variety of mineral occurrences and shows potential for mineral commodities including gold, diamonds, base metals and rare metals. During the present survey, 89 rock samples were assayed for metals including gold, silver, platinum, palladium and a range of base metals (Table 10). In addition, studies were made of gold grains in till near the Sachigo River Mine (Stone 2000b) and for kimberlite indicator minerals in a variety of surficial materials throughout the northern Superior area (Stone 2001). The results of these assays and studies are discussed with the various mineral commodities below.

GOLD

Prospectors actively searched for gold in northern greenstone belts in the 1930s. Their work led to discovery of the God's Lake Mine that operated at Gods Lake, Manitoba from 1935 to 1943 and produced 4990 kg of gold from 490 866 tonnes of ore (Richardson et al. 1996). Further prospecting led to discovery of the Sachigo River Mine at Foster Lake (see map in back pocket) and several gold occurrences in the area of Stull Lake. The Sachigo River Mine operated from 1938 to 1941 and produced 1634.8 kg of gold and 190. 5 kg of silver from 71 995 tonnes of ore in a narrow high-grade vein. Edwards (1944) provided an excellent account of the discovery, the geology and the difficulties associated with the brief mining operation at the Sachigo River Mine. Although the area of the Sachigo River Mine was explored sporadically since the cessation of mining operations (Assessment Files, Resident Geologist's Office, Red Lake), no significant new mineralization has been discovered.

Table 10. Assay data for the northern Superior area.

Sample No	Area	UTM east	UTM north	Rock Type	Assay Results								Other
					Au	Ag	Cu	Pb	Zn	Ni	Pt	Pd	
					(ppb)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppb)	(ppb)	
95DS078	S Rapson Bay	535150	6023800	10g	5						<10	<5	
95DS088	Central Stull	526500	6026200	9b	5						<10	<5	
95DS092	N Stull	524050	6030700	qcv	<3						10	<5	
95DS244	N Stull	527000	6033000	qcv,9t	19						<10	<5	
95DS245	N Stull	528400	6032800	qcv9b	32						<10	<5	
95DS168	W central Stull	524100	6028150	qcv	7						<10	<5	
95DS169	Gilleran	535300	6022100	10d	<3						<10	<5	
95DS095	Gilleran L	534500	6021400	5a	13	<20	806	231	>5000	141	<10	<5	
96DS64	Ponask Lake	523700	5984300	5a	39	3							Sb=85ppm
96DS132	Ponask Lake	529300	5979400	5a	53	3	323	55	1705	191			
96DS300	Gilleran Lake	536000	6021000	5a	5	4	453	<40	1092	157			
96DS301	Gilleran Lake	536000	6021000	5a	29	3	457	<40	536	222			
96DS303	Gilleran Lake	536000	6021000	5a			296	<40	792	145			
96DS304	Gilleran Lake	536000	6021000	5a	32	3	493	<40	246	104			
96DS305	Gilleran Lake	536000	6021000	5a	5	3	239	<40	630	143			
96DS306a	Gilleran Lake	536000	6021000	5a	<3	4	135	89	>5000	197			
96DS306b	Gilleran Lake	536000	6021000	5a	<3	4	<100	<40	>5000	181			
96DS307	Gilleran Lake	536000	6021000	5a	<3	3	402	<40	2568	119			
96DS308	Gilleran Lake	536000	6021000	5a	<3	3	134	<40	202	60			
96DS309	Gilleran Lake	536000	6021000	5a	<3	2	251	<40	445	<40			
96DS310	Gilleran Lake	536000	6021000	5a	<3	4	204	<40	389	<40			
96DS311	Gilleran Lake	536000	6021000	5a	<3	2	305	<40	638	74			
96DS313a	E. Pierce Lake	513800	5994200	7f	<3	2							
96DS313b	E. Pierce Lake	513800	5994200	7f	<3	3							
96DS314	E. Pierce Lake	513500	5995000	7f	<3	3							
97DS137	SE Pierce	505950	5996400	5a	<3	4	1100	<7	56	760	<10	<5	
97DS158	W of Rieder	561300	6083300	5a	<3	2					---	---	
98DS124	Lacey L.	580906	6036475	12m	6	<3.43					nd	nd	
98DS303	Lacey L.	581633	6040372	10a	9						8	8	
98DS127	Sellen L	589300	6040200	9s	5	<3.43					4	3	
98DS41	Thorne L	589700	6041900	9a	12	<3.43					1	nd	
98DS44	Gummer L	595100	6038900	qv	6	<3.43					1	3	
98DS49	Sellen L	589600	6040900	qv	nd	<3.43					nd	nd	
98DS50	Sellen L	588500	6040600	9a,f	1	<3.43					nd	nd	
98DS50b	Sellen L	588500	6040600	9a,f	3	<3.43					nd	nd	
98DS52	Sherman L	602500	6040400	qv	8	<3.43					1	1	
98DS55	E Foster L	606300	6041300	15a	14	<3.43					nd	nd	
98DS316	Schmidt L	610700	6040100	9v	7	<3.43					nd	nd	
98DS56c	Foster L	603100	6041800	qv	7066	<3.43					3	4	
98DS56d	Foster L	603100	6041800	qv	5331	<3.43					3	4	
98DS56e	Foster L	603100	6041800	9s	994	<3.43					9	4	
98DS56g	Foster L	603100	6041800	5a	4636	<3.43					1	nd	
98DS56h	Foster L	603100	6041800	12a	49	<3.43					nd	nd	
98DS56i	Foster L	603100	6041800	12a	172	<3.43					nd	nd	
98DS56j	Foster L	603100	6041800	qv	116230	<3.43					nd	nd	
98DS56k	Foster L	603100	6041800	qv	155269	<3.43					nd	nd	
98DS69	Stull L	531900	6023600	9b	199	<3.43					2	7	
98DS70	Stull L	531600	6024100	9b	280	<3.43					4	2	
98DS71	Gilleran	539000	6020800	qv	60	<3.43					4	3	
98DS71b	Gilleran	538600	6020900	qv	41	<3.43					2	1	
98DS71c	Gilleran	538000	6021000	5g,qv	5	<3.43					nd	nd	
98DS231	N Dodson L	625746	6068850	10g			136		86	58			
99DS60a	Tamblyn	594420	6020570	5s			190	5.17	134	56			
99DS60b	Tamblyn	594420	6020570	qv	<5	<.34					<8	<8	
99DS63	SW Tamblyn	590060	6013790	10r	<5		20		60	10	<8	<8	
99DS65	S. Meston	584480	6013890	10g			11		83	52	<8	<8	
99DS67	NE Igelstrom	587100	6008000	5s	<5	<.34					<8	<8	
99DS68a	S. Tamblyn	590100	6013000	10g	<5		27		132	15	<8	<8	
99DS68b	S. Tamblyn	590100	6013000	10g	<5		24		74	9	<8	<8	
99DS69	W. Tamblyn	587970	6014170	10g	<5		52		133	57	<8	<8	
99DS80a	N. Igelstrom	581500	6012000	qv	1170	2.06							Mo=18352 ppm
99DS80b	N. Igelstrom	581500	6012000	qv	142	0.343							Mo= 18175
99DS93	Rapson Bay	532730	6028700	9t	17	<.34	72	11.21	391	37	<8	<8	
99DS93b	Rapson Bay	532730	6028700	9t	171	1.37	280	1081.97	14337	71	<8	<8	
99DS96	Rapson Bay	531900	6029800	9mqv	1061	<.343							
99DS300	Rapson Bay	531796	6030515	9m	54	<.343					<8	<8	

Table 10 continued.

Sample No	Area	UTM east	UTM north	Rock Type	Assay Results								Other
					Au	Ag	Cu	Pb	Zn	Ni	Pt	Pd	
					(ppb)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppb)	(ppb)	
99DS304	Rapson Bay	530766	6031248	9m	116	< .34					<8	<8	
99DS305	Rapson Bay	534281	6033005	5s	8	< .34					8	8	
99DS307	Red Sucker	465000	6002000	13a	Be-3.2 ppm, Li=96.6 ppm, Rb>400 ppm, Cs=30.71 ppm, Ta =55.1 ppm								
99DS131	Matthews L.	587146	6018185	10r	<5	< .34	<5	3.74	83	<5	<8	<8	
99DS134	S. Baker L.	590974	6018933	5a	<5	< .34	19		27	<5	<8	<8	
99DS156	E. of Gilleran	538849	6020963	qv	<5	< .34					<8	<8	
99DS157	E. of Gilleran	538720	6020963	5s	<5	< .34					<8	<8	
99DS158	E. of Gilleran	538720	6020963	5s	<5	< .34					<8	<8	
99DS159	E. of Gilleran	538720	6020963	5s	<5	< .34					<8	<8	
99DS160	E. of Gilleran	538504	6021003	5a	<5	< .34					10	9	
99DS161	E. of Gilleran	538402	6021015	qv	27	< .34					<8	<8	
99DS212	Chain of Lakes	592564	6006496	10r?	<5	< .34	19	3.99	27	<5	<8	<8	
00DS78	E Blackbear	607600	6005500	10g	2.11	<3.4					7.36	7.55	
00DS88b	Rapson Bay	534265	6028195	10d	1.55	<3.4	3754	x	97		1.27	0.98	
00DS89	Rapson Bay	535578	6028969	qv	1.62	<3.4							
00DS91	Stull Lk	524704	6034088	12p	1.63	<3.4							
00DS94	Stull Lk	599950	6027800	qv	9.2	<3.4							
00DS99	Stull Lk	530684	6024911	9d	10.97	<3.4							
00DS139	N Swan	614556	6017359	9n	6.23	<3.4	109	x	55		5.22	4.49	
00DS140	N Swan	612377	6014633	9f	4.42	<3.4							
00DS148	NE Blackbear	602248	6006012	10g	3.14	<3.4	63	x	90		0.74	1.03	
00DS154	SE Schmidt Lk	617900	6035461	Qv	2.84	<3.4							
00DS205	Namaypoke	608212	6004825	10g	6.87	<3.4	127		85		0.56	0.62	

The mineralized vein at the Sachigo River Mine is unexposed at surface, however the mining operation showed that the vein trends 100° over a distance of 170 m and dips steeply to the south (Edwards 1944). It is spatially associated with *en echelon* tonalite dikes cutting mafic volcanic flows of the Ellard assemblage. Inspection of material in the mine dump represents the only opportunity to study the mineralization. The majority of rock in the mine dump is a brown-weathered fine-grained intermediate intrusive rock probably representative of the tonalite dikes that occur beside the mineralized vein. Seven samples (98DS56c to 98DS56k of Table 10) of quartz-bearing material from the mine dump yielded generally high gold tenor.

Stone (2000b) analyzed the particulate gold content of 50 samples of till collected in the area of the Sachigo River Mine. Although the survey failed to detect an anomaly of gold grains in till near and down-ice from the mineralized vein, the till sampling survey showed anomalous numbers of gold grains broadly dispersed in the area of the Sachigo River Mine. Stone (2000b) concluded that till sampling is a useful method of exploring for gold mineralization where bedrock exposure is poor, such as in the northern Superior area.

Regional mapping (Stone, Hallé and Lange 2000a) shows that the Sachigo River Mine occurs 1 km north of a small (1 km diameter) oval intrusion of the sanukitoid suite. Beakhouse (2001) noted a temporal and probable genetic relation between gold mineralization at Hemlo, Ontario and late intermediate to felsic, mantle-derived intrusive rocks such as those of the sanukitoid suite. The highly oxidized plutons are capable of generating metal-rich magmatic volatile phases that circulate through country rocks and may lead to deposition of gold in structural traps such as shear zones and veins near the intrusion. In view of this model, the Ellard Lake greenstone belt, which is intruded by a series of sanukitoid plutons between Echoing Lake and Foster Lake (see map in back pocket) may be prospective for gold.

Early prospecting led to several gold discoveries at Stull Lake including the Ellard occurrence on a small island in northern Stull Lake and the Wynne Bay South occurrence on the southeast side of the lake (Stone, Hallé and Pufahl 2001b). At the Ellard occurrence, gold is reported in association with quartz-carbonate veins in strongly deformed and carbonate-altered volcanic rocks of the Oxford Lake volcanic assemblage, although sample 95DS244 from this locality yielded only low gold tenor. At the Wynne Bay South occurrence, gold is associated with sedimentary sequences of the Oxford Lake sedimentary assemblage that are strongly sheared within the Stull–Wunnummin fault. Values of up to 280 ppb gold were obtained from samples 98DS69 and 98DS70 (*see* Table 10) from the Wynne Bay South occurrence.

In the 1980s, Westmin Resources Ltd. and Tanqueray Resources Ltd. did extensive diamond drilling at Little Stull Lake, Manitoba. Their work defined an inventory of 7800 kg of gold in 750 000 tonnes of rock from a series of zones over a 3 km strike length (Richardson et al. 1996). The mineralization occurs along a splay of the Stull–Wunnummin fault largely under the water of Little Stull Lake (see locations of showings on map in back pocket). Wolfden Resources Ltd. subsequently did further diamond drilling on the Little Stull Lake occurrence and obtained assays of up to 16.28 g/t over a 1 m length (Wolfden Resources Ltd., Press Release, September 19, 2000).

Diamond drilling by Noranda Mining and Exploration Inc. in 1989 led to discovery of the Twin Lakes gold-bearing zone (Monument Bay Occurrence) at Twin Lakes, Manitoba (see map in back pocket). The gold mineralization is associated with quartz veins cutting deformed and strongly sericitized and silica-altered volcanic and sedimentary rocks of the Oxford Lake volcanic and sedimentary assemblages. Recent diamond drilling by Wolfden Resources Ltd. in joint venture with Bema Gold Corporation has outlined a high-grade inferred resource of 13 011 kg of gold at 20.4 g/t (Bema Gold Corporation, Press Release, December 11, 2003). Regional mapping (Stone, Hallé and Pufahl 2001a,b) demonstrates that the Twin Lakes gold-bearing zone is localized along splays of the Stull–Wunnummin fault.

Other notable gold occurrences identified during the present study include an extensive quartz stockwork within faulted volcanic rocks, 1 km east of Gilleran Lake (see map in back pocket). Although low gold values were obtained from grab samples (*see* samples 98DS71 to 71c and 99DS156 to 161 of Table 10), the extent of the quartz stockwork over several hundred metres as well as its association with the Stull–Wunnummin fault suggest that this locality is worthy of further exploration. Good gold values (samples 99DS80a,b of Table 10) were also obtained from sheared granite, 10 km southwest of Tamblyn Lake (see map in back pocket and Stone, Hallé and Lange 2000b). The gold occurs with molybdenite in narrow quartz veins associated with splays of the Stull–Wunnummin fault.

In summary, the northern Superior area shows excellent potential for gold mineralization. Many gold occurrences including those at Twin Lakes and Little Stull Lake are closely associated with the various splays of the Stull–Wunnummin fault as well as alkaline, mantle-derived intrusive and extrusive rocks of the sanukitoid suite and Oxford Lake volcanic assemblage. The Stull Lake area has many similarities to the Kirkland Lake gold camp where gold mineralization is developed within a variety of strongly altered alkalic intrusive and extrusive rocks cut by the Larder Lake–Cadillac deformation zone (Thompson et al. 1950; Fyon et al. 1992; Wilkinson, Cruden and Krogh 1998). Following the model of the Kirkland Lake gold camp, gold mineralization would be expected to occur within deformed and altered plutonic and supracrustal sequences proximal to the Stull–Wunnummin fault and major splays of this fault. In particular, areas showing evidence of extensive alteration and mineral assemblages dominated by chlorite, carbonate and quartz are prospective. Although few gold occurrences are known to be associated with the South and North Kenyon faults, this may be largely due to poor exposure and little exploration. The South and North Kenyon faults as well as the Stull–Wunnummin fault are major faults of regional extent that have likely provided pathways for fluid movement and gold mineralization.

Gold exploration should be focused on these structures. Regional till-sampling surveys provide a method for identifying the most prospective parts of the area.

DIAMONDS

Although no diamonds or potentially diamondiferous host rocks have been identified, the northern Superior area contains abundant and widely distributed kimberlite indicator minerals. The kimberlite indicator minerals include pyrope garnet, picro-ilmenite, chrome diopside, chromite and olivine that are found in surficial materials (Stone 2001). During the present survey, 205 samples of beach sand, modern alluvium, till, glaciolacustrine and glaciofluvial deposits were collected and processed for kimberlite indicator minerals. Kimberlite indicators are concentrated in beach sand more than other surficial materials and occur at localities such as Stull Lake and McLeod Lake (see map in back pocket). Stone (2001) noted that the kimberlite indicator minerals have worn surfaces and have probably been transported sub-glacially from their source.

The ratio of the number of kimberlite indicators grains per 10 kg sample of beach sand is approximately 2:1 in the northern Superior area. High ratios are also found to the northwest in Manitoba (Fedikow et al. 1998; 2001a; 2001b) and to the east at Kasabonika, Ontario (Stone 2002). In contrast, the number of kimberlite indicator minerals per sample is more than an order of magnitude lower in the southern part of the western Superior Province represented by the Berens River and Sapawe areas (Stone 1994 and unpublished data). These data support the interpretation of Sage (1999) who recommended a series of prospective diamondiferous areas extending broadly along the northeast margin of Ontario including the Hudson Bay Lowlands from the Manitoba border through Attawapiskat and hence, southeast to the Kirkland Lake area. The Northern Superior Superterrane occurs within this area and represents a region of ancient crust that is a potentially favourable host for fertile kimberlites.

BASE METALS

Among the earliest recorded exploration for base metals is work done in 1960 by Phelps Dodge Corporation of Canada Limited at Stull Lake (Atkinson, Parker and Storey 1990). The work consisted of diamond drilling at the Rapson Bay–Stull Lake occurrence (*see* Pb, Zn occurrence on western Rapson Bay of Stull Lake on map in back pocket). Over the next two decades parts of the northern Superior area were explored by major exploration companies including the Canadian Nickel Company Limited, Noranda Exploration Company Limited and Sherritt Gordon Mines Limited (Assessment Files, Resident Geologist's Office, Red Lake). The recorded exploration consists mainly of logs of widely scattered diamond-drill holes and a few ground geophysical surveys. It can be assumed however that the diamond drilling must have been preceded by airborne geophysical surveys that were used to select diamond-drill targets but were not recorded for assessment purposes. Areas of work included parts of the Stull–Swan lakes greenstone belt at Stull Lake, Gilleran Lake, Joe Lake, Wintering Lake, Blackbear Lake and Swan Lake (see map in back pocket). The Ellard Lake greenstone belt was explored in the area between Ellard Lake and Echoing Lake. The Ponask Lake greenstone belt was explored for base metals and asbestos at western Ponask Lake.

Little base metal exploration is recorded in the northern Superior area since the 1980s although the results of the present study and companion studies in Manitoba indicate several areas of significant base metal potential. These are outlined briefly below.

A variety of mineralogic and geochemical data suggests that the Ponask Lake greenstone belt may be prospective for Cu-Ni and volcanogenic massive sulphide mineralization. For example, samples of surficial materials in the Ponask Lake area contain anomalous numbers of chromite grains interpreted to have been derived locally from komatiite sequences within the Ponask Lake greenstone belt (Stone 2001). Polished thin sections of komatiite at Ponask and Sachigo lakes contain chromite as well as finely disseminated chalcopyrite and pentlandite. Also of note are samples 97DS137 and 96DS132 (*see* Table 10) from rusty zones within mafic volcanic and gabbro sequences at Ponask Lake and eastern Pierce Lake, respectively. Sample 97DS137 has anomalous Cu and Ni characteristic of base metal mineralization associated with mafic to ultramafic sequences.

Sample 96DS70 (*see* Figure 8c and file *North Superior Geochemistry 2000* on MRD 135), which represents the oval felsic volcanic domain at the west end of Ponask Lake, has the chemical characteristics of an FIII rhyolite of Lesher et al. (1986). FIII rhyolites have been interpreted to represent high-level magma chambers capable of driving hydrothermal cells with possible associated volcanogenic massive sulphide mineralization. Accordingly, the western Ponask Lake area may be prospective for volcanogenic massive sulphide mineralization. Sample 96DS132, which is Zn-rich, may represent an occurrence of this type of mineralization at Ponask Lake.

Several base metal occurrences are identified in the Stull–Gilleran lakes area. The Rapson Bay–Stull Lake occurrence consists of seams (up to 1 m wide) of rusty disseminated sulphide minerals in felsic volcanic sequences exposed in shoreline outcrops. Samples 99DS93 and 99DS93b from this occurrence contain anomalous Zn, Pb and Cu (*see* Table 10). Sample 00DS88 was taken from the Rapson Bay chalcopyrite occurrence on an island about 1 km east of the Rapson Bay–Stull Lake occurrence[7]. This sample contains 0.37% Cu, which is derived from chalcopyrite occurring as fracture fillings and disseminated blebs within diorite of the Gilleran Lake intrusion. Disseminated chalcopyrite is also noted in thin sections of diorite and gabbro elsewhere within the Gilleran Lake intrusion such as on islands in Gilleran Lake.

Zn mineralization is noted in mafic sequences on the south shore of Gilleran Lake. At low water conditions, local alteration zones containing rusty seams up to 0.1 m wide can be observed in outcrops of mafic volcanic rocks at the shoreline. The alteration zones contain a mineral assemblage dominated by chlorite+zoisite+feldspar+garnet+staurolite (*see* sample 96DS300 in files *Samples North Superior 2004*, *Garnet* and *Staurolite* on MRD 135). The rusty seams contain disseminated sphalerite within the above assemblage of silicate minerals and assays of this material yielded anomalous Zn, Cu and Ni (*see* samples 95DS095 and 96DS300-311 of Table 10).

In summary, the majority of base metal occurrences in the Stull Lake area appear to be concentrated within and at the margins of the Gilleran Lake intrusion. This late, mantle-derived alkaline intrusion may have been a source for metals or acted as a heat-engine driving convective hydrothermal cells for the redistribution of metals within country rocks. Further exploration for base metals should be focused in the area of the Gilleran Lake intrusion.

Corkery and Skulski (1998) noted a broad alteration zone extending through mafic volcanic sequences south of Little Stull Lake, Manitoba. Alteration appears to be controlled by fractures within the zone and is characterized by bleaching of basalt and gabbro with local development of white silica-rich zones and gossan zones. The survey of Fedikow et al. (1998) demonstrated anomalous Cu, Co and As within soils overlying this zone extending south from Little Stull Lake to the Ontario border northeast of Stull Lake. The alteration zone represents another target for base metal exploration in the Stull Lake area.

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[7] Locations of the Rapson Bay-Stull Lake and Rapson Bay chalcopyrite occurrences are shown on Stone, Hallé and Pufahl, 2001a.

Rusty zones were observed at scattered localities in all greenstone belts of the northern Superior area, however assays showed no significant concentrations of base metals. The gabbro body 5 kilometres southwest of Tamblyn Lake was sampled at several localities and assayed for base metals and platinum-group metals. The results for samples 99DS63-69 of Table 10 show negligible content of these metals.

RARE METALS

Rare metals include Li, Rb, Cs, Be, Nb, Ta and Ga and are typically associated with minerals such as spodumene, lepidolite, beryl, and columbite-tantalite in highly fractionated, commonly dike-like phases of the peraluminous granite suite. Although no significant occurrences of rare metals were identified in the present area, several are known in adjacent parts of Manitoba. For example, Chackowsky et al. (1985) documented strongly deformed pegmatite dikes within mylonitic basalt sequences at Red Cross Lake, Manitoba. The deformed pegmatites contain Rb-rich lepidolite and Rb-rich potassium feldspar and occur 50 km west northwest of the Ontario border along the North Kenyon fault.

Spodumene-bearing pegmatites are noted at Johnson Bay of Gods Lake (Chakowsky et al. 1985) and at Cross Lake, Manitoba (Lenton and Anderson 1983). The Cross Lake pegmatite field is also mineralized with columbite-tantalite. Lenton (1985) mapped a pluton of coarse-grained white granite at Magill Lake (south of Knee Lake, Manitoba). The Magill Lake pluton intrudes sedimentary rocks of the Oxford Lake sedimentary assemblage and contains accessory garnet, tourmaline and apatite; it probably belongs with the peraluminous granite suite. Lenton (1985) noted spodumene-bearing pegmatite dikes at McLaughlin Lake, which is situated at the faulted south margin of the Oxford Lake sedimentary sequence containing the Magill Lake pluton.

Chackowsky and Cerny (1984) and Gunter and Chakowsky (1988) documented a variety of rare-metal mineralization at Red Sucker Lake, Manitoba. In this area, three types of pegmatitic leucogranite dikes are recognized within migmatitic supracrustal rocks. These include dikes mineralized with albite and spodumene, cassiterite and petalite. Red Sucker Lake occurs approximately 40 km west of Pierce Lake and represents the western extension of the Ponask Lake greenstone belt.

The majority of rare metal occurrences in northern Manitoba are within highly fractionated pegmatite dikes of the peraluminous granite suite. Many dikes cut migmatized sedimentary sequences and are localized along major faults such as western extensions of the Stull–Wunnummin fault and North and South Kenyon faults. Within the present area, peraluminous granite occurs as small units associated with sedimentary migmatites in the Pierce Lake area. Peraluminous granite occurs in a unit about 10 km north of Joe Lake (see map in back pocket) and as dikes at scattered localities in all greenstone belts. Further exploration for rare metal pegmatites should be directed at highly metamorphosed sedimentary units, particularly where they are cut by major faults.

The carbonatite pluton ("Carb" Lake carbonatite), 10 km north of McLeod Lake, represents a potential source for rare metal mineralization. Although unexposed, the carbonatite pluton was diamond drilled by Big Nama Creek Mines Ltd. in 1968 and tested for niobium (Assessment Files, Resident Geologist's Office, Red Lake). As part of the assessment of sources of heavy minerals in surficial materials, Stone (2001) produced heavy mineral concentrates from samples of the diamond-drill core stored in the Drill Core Library at Kenora, Ontario and obtained trace quantities of pyrochlore and synchysite. Sage (1987) recommended that the "Carb" Lake carbonatite be explored for surface depressions or solution cavities where residual accumulations of apatite, vermiculite, pyrochlore and minerals containing rare earth elements can be concentrated.

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Metric Conversion Table

Conversion from SI to Imperial			Conversion from Imperial to SI		
SI Unit	*Multiplied by*	*Gives*	*Imperial Unit*	*Multiplied by*	*Gives*
		LENGTH			
1 mm	0.039 37	inches	1 inch	**25.4**	mm
1 cm	0.393 70	inches	1 inch	**2.54**	cm
1 m	3.280 84	feet	1 foot	**0.304 8**	m
1 m	0.049 709	chains	1 chain	20.116 8	m
1 km	0.621 371	miles (statute)	1 mile (statute)	**1.609 344**	km
		AREA			
1 cm^2	0.155 0	square inches	1 square inch	**6.451 6**	cm^2
1 m^2	10.763 9	square feet	1 square foot	**0.092 903 04**	m^2
1 km^2	0.386 10	square miles	1 square mile	2.589 988	km^2
1 ha	2.471 054	acres	1 acre	0.404 685 6	ha
		VOLUME			
1 cm^3	0.061 023	cubic inches	1 cubic inch	**16.387 064**	cm^3
1 m^3	35.314 7	cubic feet	1 cubic foot	0.028 316 85	m^3
1 m^3	1.307 951	cubic yards	1 cubic yard	0.764 554 86	m^3
		CAPACITY			
1 L	1.759 755	pints	1 pint	0.568 261	L
1 L	0.879 877	quarts	1 quart	1.136 522	L
1 L	0.219 969	gallons	1 gallon	**4.546 090**	L
		MASS			
1 g	0.035 273 962	ounces (avdp)	1 ounce (avdp)	28.349 523	g
1 g	0.032 150 747	ounces (troy)	1 ounce (troy)	**31.103 476 8**	g
1 kg	2.204 622 6	pounds (avdp)	1 pound (avdp)	**0.453 592 37**	kg
1 kg	0.001 102 3	tons (short)	1 ton (short)	**907.184 74**	kg
1 t	1.102 311 3	tons (short)	1 ton (short)	**0.907 184 74**	t
1 kg	0.000 984 21	tons (long)	1 ton (long)	**1016.046 908 8**	kg
1 t	0.984 206 5	tons (long)	1 ton (long)	**1.016 046 90**	t
		CONCENTRATION			
1 g/t	0.029 166 6	ounce (troy)/ ton (short)	1 ounce (troy)/ ton (short)	34.285 714 2	g/t
1 g/t	0.583 333 33	pennyweights/ ton (short)	1 pennyweight/ ton (short)	1.714 285 7	g/t

OTHER USEFUL CONVERSION FACTORS

	Multiplied by	
1 ounce (troy) per ton (short)	31.103 477	grams per ton (short)
1 gram per ton (short)	0.032 151	ounces (troy) per ton (short)
1 ounce (troy) per ton (short)	20.0	pennyweights per ton (short)
1 pennyweight per ton (short)	0.05	ounces (troy) per ton (short)

Note: Conversion factors which are in bold type are exact. The conversion factors have been taken from or have been derived from factors given in the Metric Practice Guide for the Canadian Mining and Metallurgical Industries, published by the Mining Association of Canada in co-operation with the Coal Association of Canada.

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